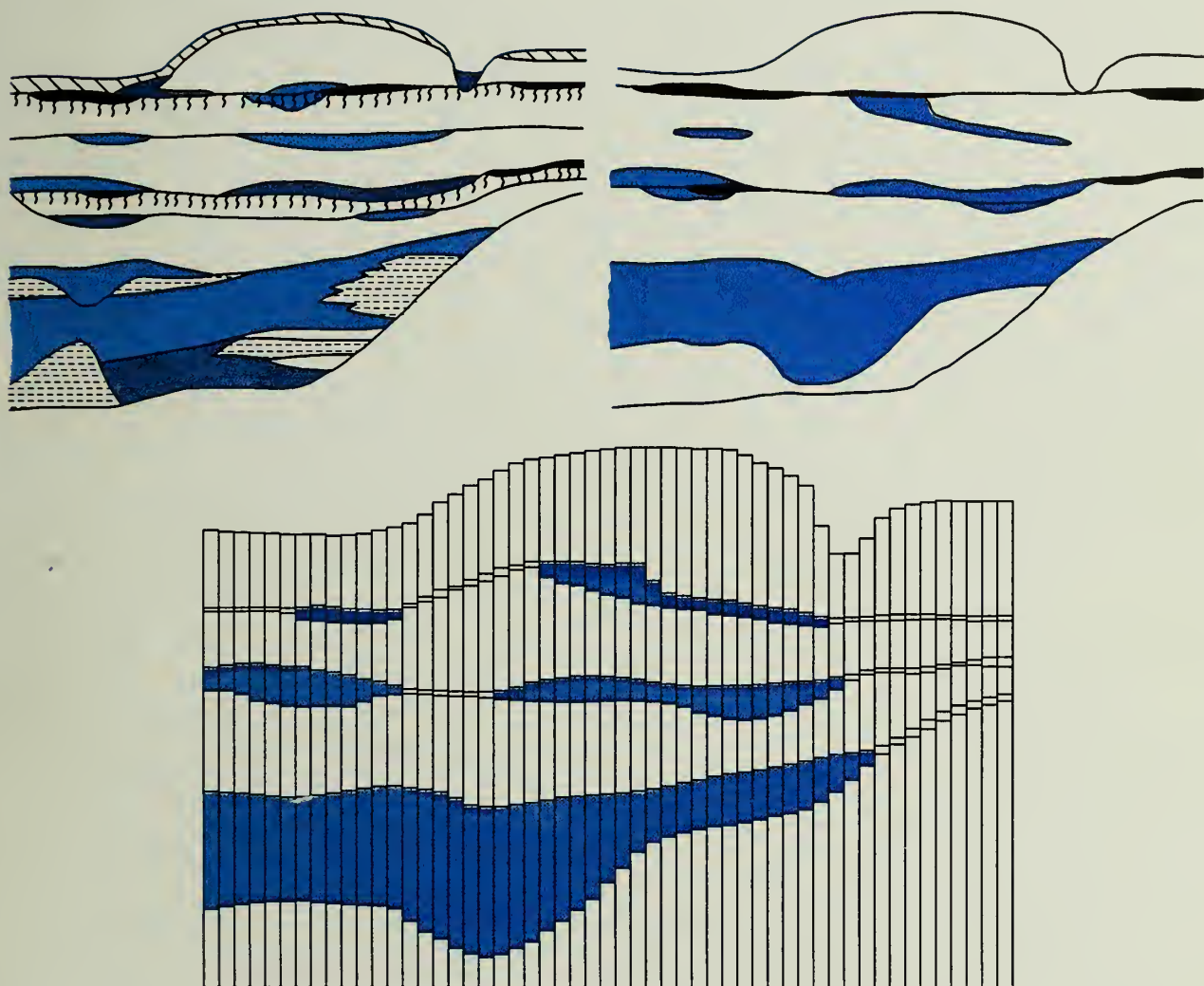


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Hydrogeology and Ground-Water Availability in Southwest McLean and Southeast Tazewell Counties

Part 2: Aquifer Modeling and Final Report



1998
Cooperative Ground-Water Report 19

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Hydrogeology and Ground-Water Availability in Southwest McLean and Southeast Tazewell Counties

Part 2: Aquifer Modeling and Final Report

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Title: Hydrogeology and Ground-Water Availability in Southwest McLean and Southeast Tazewell Counties, Part 2: Aquifer Modeling and Final Report

Abstract: An assessment of the aquifer characteristics and ground-water availability has been completed for one of the largest sand-and-gravel aquifers in Illinois, the Sankoty-Mahomet Sand aquifer, in the confluence area of the Mahomet and Mackinaw Bedrock Valleys in McLean and Tazewell Counties. Based on data gathered during the field portion of the study, a ground-water flow model was developed that simulates the effects of a hypothetical well field for various locations within the study area. It was used to estimate the quantity of ground water that could be withdrawn from the aquifer within the study area. The model was calibrated to match water levels measured in area wells in 1994 and to match the baseflow gains and losses in the Mackinaw River and Sugar Creek. A hypothetical well field, pumping 15 million gallons a day (mgd), was simulated at four locations. The results varied from a maximum drawdown of 8 feet in the Hopedale scenario to 55 feet in the Armington scenario. If a well field similar to the well fields modeled were installed in the study area, as many as 400 private wells may be impacted. In certain areas near the Mackinaw River, a well field would greatly reduce the ground-water portion of baseflow entering the Mackinaw River. Pumping three of the well fields together, at a total rate of 37.5 mgd, indicated that the aquifer should be able to sustain withdrawals in excess of 37.5 mgd if the pumpage is distributed in the study area.

Reference: Wilson, S.D., G.S. Roadcap, B.L. Herzog, D.R. Larson, and D. Winstanley. *Hydrogeology and Ground-Water Availability in Southwest McLean and Southeast Tazewell Counties, Part 2: Aquifer Modeling and Final Report*. Illinois State Water Survey and Illinois State Geological Survey Cooperative Ground-Water Report 19, Champaign, IL, 1998.

Indexing Terms: aquifer assessment, ground-water availability, ground-water flow model, hydrogeologic assessment, Modflow, regional well field, Sankoty-Mahomet Sand aquifer, water supply

Cover illustrations: Top left, representative cross section depicting the geology of the study area. Top right, representative hydrogeologic cross section of the study area depicting the simplifications made to the geologic interpretations for input into the ground-water flow model. Bottom, converted representative hydrogeologic cross section of the study area as utilized in the ground-water flow model. Aquifers are blue.

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PREFACE

Over the past million years or more, great climatic shifts have resulted in repeated expansion and recession of glaciers in large parts of the Northern Hemisphere, including most of Illinois. These glacial movements left behind a collection of sediments over the bedrock, and Illinois reaps many benefits from these glacial deposits. Soils on top of these mineral-rich glacial sediments are so fertile and productive that Illinois provides approximately 6 percent of the world's corn (16 percent of the U.S. corn) and 8 percent of the world's soybeans (17 percent of the U.S. soybeans).

Sand-and-gravel aquifers beneath the land surface contain vast quantities of ground water. Glacier-deposited, clay-rich, relatively impermeable glacial tills may enclose this water supply in one or more irregularly shaped layers of porous, permeable sand and gravel. Tens of thousands of wells drilled into these aquifers throughout the state provide reliable sources of water for many Illinoisans.

As demand for water increases, an important challenge is to better characterize the water-bearing potential of our glacial-drift aquifers so these ground-water resources can be managed prudently. Over the past 20 years, many surface water supplies have experienced shortages—such as those that developed during the 1988 drought in the Midwest—and surface and ground-water supplies also have come under scrutiny because of water-quality concerns.

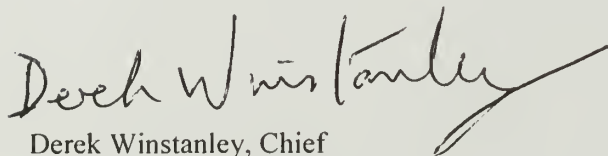
The "crown jewel" of central Illinois aquifers is the Sankoty-Mahomet Sand aquifer. In southwest McLean and southeast Tazewell Counties, this aquifer lies buried several hundred feet below the land surface in preglacial bedrock river valleys. This aquifer is as much as 150 feet thick in places and is capable of a sustained yield of millions of gallons of water per day, enough to provide a reliable supply of water to meet the increasing demands of many central Illinois communities in the 21st century. However, to ensure protection of the Sankoty-Mahomet Sand aquifer from overuse and contamination, it is necessary to understand the complex hydrogeological setting of this aquifer, the ground-water flow system within it, and its relationship to the overlying aquifers.

The Illinois State Water Survey (ISWS) and the Illinois State Geological Survey (ISGS) are proud to have been selected by the Long Range Water Plan Steering Committee (LRWPSC) to conduct studies to evaluate the ground-water resources of southwest McLean and southeast Tazewell Counties. Survey scientists have responded to the project's many challenges by applying state-of-the-art digital computer modeling and geophysical exploration techniques.

This report describes the large databases and analytical tools and methods that have been used to provide answers to questions posed by the LRWPSC. However, the main product of the project is a mathematical, computer-based ground-water flow model. The conclusions presented in this report are drawn from the results of

modeling scenarios. Some advantages of the model are that it can be updated with new data, and it is flexible enough to simulate any number of water-withdrawal scenarios that the LRWPSC may wish to investigate. This state-of-the-art planning tool establishes the ISWS, the ISGS, and the communities of central Illinois as leaders in the science of ground-water resource evaluation. However, further monitoring and research need to be conducted in support of the wise and effective management of Illinois' precious ground-water resources in the decades ahead.

On behalf of the Illinois Department of Natural Resources and the Scientific Surveys, we are pleased to present this report to the LRWPSC. We look forward to future opportunities to work with your communities in evaluating natural resources through the use of established and innovative science and engineering techniques.



Derek Winstanley, Chief
Illinois State Water Survey



William W. Shilts, Chief
Illinois State Geological Survey

Illinois Department of Natural Resources
Office of Scientific Research and Analysis

EXECUTIVE SUMMARY

In 1993, with funding from the Long Range Water Plan Steering Committee (LRWPSC), the Illinois State Water Survey (ISWS) and the Illinois State Geological Survey (ISGS) began a study of the sand-and-gravel aquifers in southwest McLean and southeast Tazewell Counties to estimate the availability of ground water and determine the hydrogeologic feasibility of possibly developing a regional water supply. The study area includes the confluence of the buried Mahomet and Mackinaw Bedrock Valleys (confluence area) and contains part of one of the largest sand-and-gravel aquifers in Illinois, the Sankoty-Mahomet Sand aquifer. The study had two goals: (1) to determine the quantity of water a well field in the Sankoty-Mahomet Sand aquifer could yield, and (2) to determine the possible impacts to ground-water levels and existing wells that might occur in the Sankoty-Mahomet Sand aquifer and overlying aquifers from the development of a well field pumping 10-15 million gallons of water a day (mgd).

Two major tasks were completed to meet the study goals. The first task was a hydrogeologic characterization of the glacially deposited (glacial-drift) aquifers within the confluence area. Results of the hydrogeologic characterization were published in 1995 (Herzog et al., 1995a and b) and a summary of their findings are in the appendices. The second task, and the subject of this report, was the development of a computer-based, mathematical model of the ground-water flow in the glacial deposits (ground-water flow model). The ground-water flow model was used to simulate the effects of a hypothetical well field for various locations within the study area and to provide an estimate of the amount of ground water a regional well field could yield from the Sankoty-Mahomet Sand aquifer within the confluence area.

The characterization of the hydrogeology of the glacial-drift aquifer system was simplified to allow the development of a ground-water flow model. The generalized hydrogeology resembled a layer cake with uneven layers, some of which were discontinuous. The layers included relatively impermeable bedrock overlain by three sand-and-gravel aquifer layers that are generally separated by aquitard layers. Due to the complexity of the spatial distribution of the sand-and-gravel deposits above the Sankoty-Mahomet Sand aquifer, these shallower deposits were generalized as two aquifer layers. Although none of the aquifer deposits represented by the shallower aquifer layers are capable of sustaining a 10-15 mgd water supply, the thickness, distribution, and hydraulic properties of these deposits are important for a complete understanding of the hydrology of the model area. In some parts of the area covered by the ground-water flow model, two or more of the aquifer layers are in direct contact, providing a "window" of hydraulic connection between the aquifer layers. In other parts of the model area, one or more of the aquifer layers are absent.

Using the information from the hydrogeologic mapping and water-level data, chloride concentrations, and percent modern carbon data from observation wells, an updated conceptual understanding of the ground-water flow system for the Sankoty-Mahomet Sand aquifer was developed that described the movement of ground water into and out of the model area. Ground water in the Sankoty-Mahomet Sand aquifer generally flows through the Mahomet Bedrock Valley from the southeast, westward to the Illinois River and northward through the Mackinaw Bedrock Valley. The natural ground-water discharge areas for the Sankoty-Mahomet Sand aquifer in the study area are the Mackinaw River in the west-central part of the study area and Sugar Creek in the southwestern part of the study area. In some areas very close to the rivers, ground water is flowing upward from the Sankoty-Mahomet Sand aquifer through the upper aquifers and into the stream beds. There is a slight hydraulic gradient (slope) east of the model area that steepens where the flow enters the study area, even though the aquifer volume increases. This slope increase is caused by a greater amount of recharge entering the aquifer due to hydraulic connections with overlying aquifers. The areas of connection are more numerous in the west and north portions of the model area, as demonstrated by increases in water levels, decreases in chloride concentrations, and increases in modern carbon isotope concentrations in the Sankoty-Mahomet Sand aquifer. Down gradient of these connections, the chloride concentrations remain low, which suggests that the influx of

ground water through these connections provides the majority of the recharge in these areas. Water pumped from the Normal west well field south of Danvers, which has wells penetrating into one of these upper aquifer connections, has low chloride values, indicative of water coming from the upper sands.

Although the size of the original study area was about 260 square miles, the area to be modeled (model area) was expanded to 1,100 square miles. This expansion was necessary to reduce the effects of the model boundary conditions on simulated water levels in the study area. The simulated water levels are strongly influenced by the boundary conditions, which reduce the accuracy of the simulated water levels near the boundaries.

The ground-water flow model was developed using Visual Modflow software. Three aquifer layers sandwiched between four aquitard layers are simulated in the model. Bedrock forms the lowest aquitard; till units form the others. The hydraulic property values of each hydrogeologic unit were assigned to the corresponding layer in the ground-water flow model where the unit was present. When a unit was absent, the layer was assigned the value of an overlying or underlying hydrogeologic layer. The model's boundary conditions control the regional flow into and out of the study area, discharge to and from the streams, infiltration from precipitation, and removal of water by production wells. The model was calibrated to match water levels measured in area wells in 1994 and to match the baseflow gains and losses in the Mackinaw River and Sugar Creek. The mean absolute error of the simulated water levels was 4.99 feet, which was only slightly greater than the errors associated with the potentiometric surface maps, indicating a good match between the model and the characterization of the ground-water flow system. The ground-water flow budget calculated using the model shows that 80 percent of the water coming into the model is from infiltration of precipitation, 11 percent is from the regional Mahomet aquifer in the east, and 8 percent is from river leakage. The budget also shows that 57 percent of the surface and ground water that leaves the model area does so through discharge to the rivers, 33 percent to the regional ground-water flow to the north and to the west, and the remaining 10 percent to existing production wells.

To meet the project goals, a hypothetical regional well field in the Sankoty-Mahomet Sand aquifer pumping 15 mgd was simulated using the model. Four locations for the well field were tested near the towns of Armington, Emden, Mackinaw, and Hopedale. Many other scenarios could have been evaluated. The results varied dramatically among the four scenarios with a maximum steady-state drawdown of 8 feet at the scenario near Hopedale and a maximum drawdown of 55 feet at the simulation near Armington. The magnitude of drawdown 2 miles away from the well fields ranged from 5 feet at the Hopedale scenario to 40 feet at the Armington scenario. Based on existing well records and using 10 feet of drawdown as the minimum criteria for requiring review, the number of private wells that would require further review to determine the potential for impacts from pumpage ranged from no wells in the Hopedale scenario to 400 wells in the Armington scenario. An additional scenario with three of the well fields together pumping 37.5 mgd was simulated to evaluate the Sankoty-Mahomet Sand aquifer response to this amount of pumpage. This scenario, though unrealistic in practice because of costs due to the distance between the sets of wells, indicated that the aquifer can sustain at least 37.5 mgd of pumpage if withdrawals were distributed in the model area.

Because of the complexity of the hydrogeology and ground-water flow system, the potential yield of the Sankoty-Mahomet Sand aquifer is highly dependent on placement of the well field. In addition, placement of a well field may affect others parts of the hydrologic system. For instance, placement of the well field near the Mackinaw River in an area of direct hydraulic connection will cause a diversion of surface water from the river into the aquifer, reducing the downstream flow of the river. Some locations, therefore, are more suitable for development of a regional water supply than others. Placing the well field at a location in which the aquifer thickness is maximized, there is connection with shallow aquifers, and aquifer boundary effects could be minimized will minimize the impacts from pumpage of a regional well field on water levels and on private water wells and provide the best potential yield in the study area.

Though the calibrated model matches the data well, it is a simplification of actual conditions because the sand-and-gravel deposits were simplified into three aquifer layers and the hydraulic data for every model cell is not known. As additional hydrogeologic and hydraulic data become available, interpretation of the hydrogeology of the aquifers may change, warranting modification of the model parameters to improve the model.

ACKNOWLEDGMENTS

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Several staff members from the Illinois State Geological Survey and the Illinois State Water Survey participated in this phase of the project. John F. Blomberg and Alison B. Lecouris helped develop the project database. Jane A. Duncan, Sandie E. Osterbur, and Deanna G. Overcast provided copies of well records and helped resolve differences between ISGS and ISWS well records. Edward C. Smith, Richard J. Rice, Brett T. Bogge, and Patricia A. Carrera helped verify well locations and determine well elevations. Richard C. Berg, E. Donald McKay, Scott C. Meyer, and William W. Shilts reviewed the geologic interpretations used in the model. Curt Abert developed the hydrogeologic layers using EarthVision©. Adrian P. Visocky suggested test locations and pumping scenarios for the hypothetical well field. Manoutcher Heidari provided assistance in the modeling effort. Richard C. Berg, Manoutcher Heidari, E. Donald McKay, Ellis W. Sanderson, William W. Shilts, Adrian P. Visocky, and H. Allen Wehrmann served as technical reviewers for the manuscript. Sean V. Sinclair, Curt Abert, Linda J. Hascall, and Jacquelyn L. Hannah produced the figures. Agnes E. Dillon and Eva C. Kingston were the technical editors.

The interpretations and conclusions expressed herein are those of the authors and do not necessarily reflect the views of the sponsor.

INTRODUCTION

Background

The 1988-1989 drought in central Illinois focused attention on the need for a reliable, long-term water supply for the Bloomington-Normal area. Little planning had been done prior to the drought for a regional water supply to meet increasing demand and to provide for drought emergencies. During the drought, however, the water level in the city of Bloomington's two reservoirs (Lake Bloomington and Evergreen Lake) dropped to alarmingly low levels, prompting Bloomington to institute severe water-use restrictions and to purchase supplemental water from the town of Normal. To solve its short-term problem, Bloomington investigated two additional water sources: (1) development of a supplemental ground-water supply from the area near its two reservoirs, and (2) withdrawal of surface water from the Mackinaw River (Farnsworth and Wylie and Hanson Engineers, 1989). A surface geophysical study conducted by the Illinois State Geological Survey (ISGS) indicated that insufficient ground water was available to provide a supplementary municipal supply for Bloomington (Larson and Poole, 1989). As an emergency measure, a side-channel pumping pool was built along the Mackinaw River in 1989. Shortly thereafter, water was pumped from it into Evergreen Lake. This pumping pool remains in place for use only during water-supply emergencies.

After the drought, officials from the city of Bloomington, the town of Normal, and McLean County formed a Joint Steering Committee to investigate the feasibility of developing a regional water system to serve west-central McLean and eastern Tazewell Counties. A study conducted for the Joint Steering Committee concluded that a regional supply would be both feasible and potentially beneficial to many communities (Farnsworth and Wylie, 1990). Four potential sources were identified for the regional water supply: (1) a reservoir to be constructed on Panther Creek, (2) a new pipeline to the Illinois River, (3) a new water-recycling system, and (4) ground water to be pumped from the Sankoty-Mahomet Sand aquifer in southwestern McLean and southeastern Tazewell Counties. Ground water from the Sankoty-Mahomet Sand aquifer was determined to be the most cost-effective source for a regional water supply (Farnsworth and Wylie, 1990).

The Long Range Water Plan Steering Committee (LRWPSC) was formed in 1992 by intergovernmental agreement among 18 public water suppliers and 4 water authorities. In 1993, the LRWPSC commissioned two studies. The first was a review of each community's water-supply needs; the second was a hydrogeologic investigation of the water-supply potential from the Sankoty-Mahomet Sand aquifer. A report evaluating the existing and future water needs for 18 community water supplies in McLean and Tazewell Counties was completed in 1994 (Farnsworth and Wylie, 1994); this report concluded that the rural communities in the aquifer characterization study area have sufficient water resources now and for the foreseeable future. Bloomington and Normal were identified as the first communities that would require water-supply expansion.

The ISGS and the Illinois State Water Survey (ISWS) conducted a hydrogeologic investigation to evaluate aquifer characteristics and ground-water availability (the current study). The study had two goals: (1) to determine the quantity of ground water a regional well field could yield, and (2) to determine the possible impacts to ground-water levels and to existing wells that might occur from the development of a 10-15 million gallons per day (mgd) well field. To meet the projected needs determined by the LRWPSC, any potential location should provide from 10-15 mgd of water while minimizing impacts to existing wells. Two major tasks were required to meet the study goals: (1) hydrogeological characterization of the aquifer system, including the aquifers and their confining units; and (2) development of a computer-based, mathematical model of the ground-water flow system (ground-water flow model).

Regional Setting

The primary research area studied and presented by Herzog et al. (1995a and b) included about 260 square miles in central Illinois (figure 1). A summary of Herzog et al. (1995a and b) is provided in appendix A. The area lies mostly in McLean and Tazewell Counties but includes part of northern Logan County. This 260-square-mile area was selected by the LRWPSC, using the results of a study by Kempton and Visocky (1992) of a 30-township area that included most of western McLean County, northern Logan and DeWitt Counties, and southeastern Tazewell County. Their work indicated that the greatest potential for obtaining a large ground-water supply is within the northwest quarter of the area they studied (figure 2), where the aquifer was thickest.

The area used for the ground-water flow model includes a border area surrounding the study area. Together these areas encompass 1,100 square miles in central Illinois (figure 1). The "model area" includes most of Tazewell County, the southwest corner of Woodford County, western McLean County, northwest DeWitt County, and approximately the northern half of Logan County. The model area is largely rural, but it includes the growing suburban areas around Bloomington, Normal, and Lincoln, as well as the Morton, East Peoria, and Pekin areas. Nineteen smaller communities are located in the model area.

Purpose and Scope

This report describes the results of the second study task, which is the development of the ground-water flow model, and it presents the conclusions of the study. To meet the study goals, a ground-water flow model was developed to simulate the ground-water flow system so reliable estimates of hypothetical water-level changes in existing wells in the study area and changes in the hydraulic head of the aquifer layers could be made.

This report guides the reader through: (1) the steps taken to develop the input data for the ground-water flow model, (2) the model calibration process, (3) the use of the model to evaluate the effects of ground-water pumpage on the aquifers and existing wells, and (4) the research conclusions and recommendations for further study. A glossary of terms is included. The following three sections of this report, Model Development, Hydrogeology, and Hydrology, explain what data were needed, collected, and evaluated for input into the ground-water flow model. The Mathematical Ground-Water Flow Model section describes how the model was developed, the data used in the model, the model calibration, and the results of the model runs at four modeled well fields. The last section, Conclusions and Recommendations, provides the conclusions of the study, provides recommendations for further study, and comment about potential additional influences on the aquifer system; it also suggests several tasks that should be completed if the planning for a regional well field moves forward.

MODEL DEVELOPMENT

This section describes the selection of the model boundaries, an overview of the methods used to conduct this study, and the quality assurance measures implemented to ensure a reasonable degree of confidence in the findings and conclusions.

Model Area

The purpose of the ground-water flow model was to simulate the flow system and changes in water levels due to withdrawal of ground water by a well field in order to evaluate the impacts of these changes on the aquifers and existing wells in the 260-square-mile study area. To accomplish this, the boundaries of the model

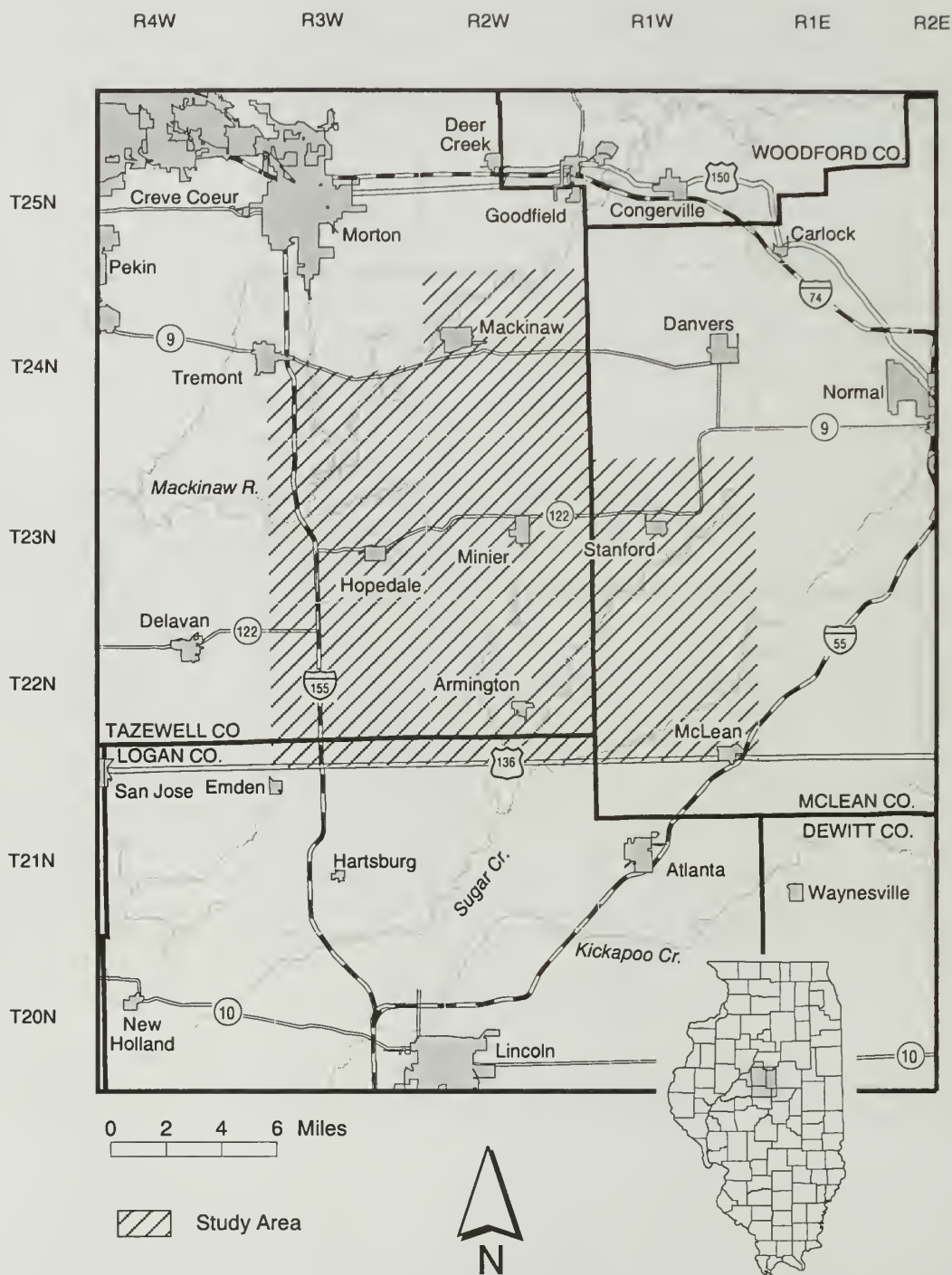


Figure 1. The model area, including the study area (cross hatched)

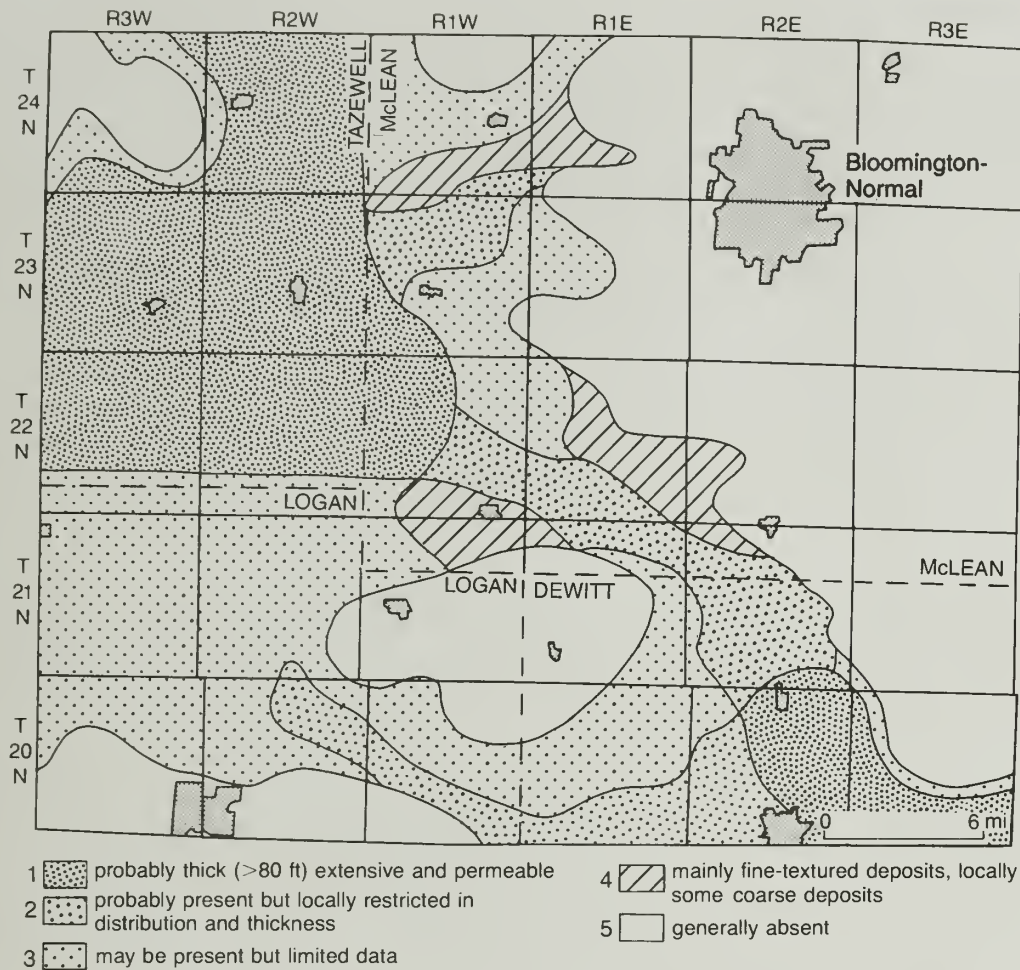


Figure 2. Occurrence of the Sankoty-Mahomet Sand aquifer
(adapted from Kempton and Visocky, 1992)

were selected so they would not influence the changes in simulated water levels within the study area. Therefore, model boundaries were selected based on the distribution of the Sankoty-Mahomet Sand aquifer because Herzog et al. (1995a) identified this aquifer as having the greatest potential for development of the desired water supply. The shallower aquifers are critical parts of the ground-water flow system and are used for domestic water supplies, but their potential for development as the source of a large ground-water supply (10-15 mgd) is minimal.

For a ground-water flow model, aquifer boundaries are defined in two ways: (1) boundaries can be set at the physical limit of an aquifer (a geologic or hydraulic feature of the aquifer), or (2) they can be set far enough from the edge of an area that pumping does not cause a noticeable change in water levels at or beyond the edge of the area. Because the Sankoty-Mahomet Sand aquifer was deposited in channels of the Mackinaw and Mahomet Bedrock Valleys, and the bedrock walls of these valleys are of low hydraulic conductivity, the walls limit the extent of the aquifer. The valley walls provide the ground-water flow model with physical boundaries in the south, northeast, and northwest parts of the model area. Where physical boundaries did not exist within the model area, the boundaries were set a minimum of 6 miles from the edge of the study area to minimize the influence of the boundary on water levels within the study area.

Overview of the Modeling Process

The process of developing the ground-water flow model involved several steps, as depicted in figure 3. The input requirements of the ground-water flow model were first determined. The location and extent of the aquifer, the location of the model boundaries, and the number of aquifers to be modeled were determined based on the hydrogeologic understanding of the area. The determination of the area modeled was partially based on the need to add a border around the study area to minimize boundary effects in the model. Then all available data pertaining to geology and hydrology (hydrogeologic data) were compiled and reviewed. Both the data obtained in the study area during the aquifer characterization study and existing data for the 840 square miles of the border area outside the aquifer characterization study area (figure 1) were included. A review of existing information revealed that additional field data would be beneficial. In one critical area, the thickness of the Sankoty-Mahomet Sand aquifer was in question. To provide an accurate thickness measurement at this location, a test hole was drilled and completed as an observation well. The beds of the major streams were sampled to identify those reaches with coarse-grained sediments below the stream bed that would likely allow for a stream/aquifer connection. All available data were entered into a hydrogeologic or ground-water-level database. These data are discussed in the section describing model inputs.

From the hydrogeologic database, a three-dimensional hydrogeologic representation consisting of seven layers (three aquifer layers, three aquitard layers, and bedrock) was developed using the Geographical Information System (GIS) software package ARC/INFO (ESRI, Redlands, CA) and the geologic modeling software EarthVision© (Dynamic Graphics Inc., Alameda, CA), as described in the section, Hydrogeologic Layer Development. The hydrogeologic representation in seven layers is a simplification of the known hydrogeology of the area. The existing data are not comprehensive enough to depict every detail of the hydrogeology in three dimensions. Even if such a complex representation could be developed, simplifications often must be made so mathematical solutions can be calculated digitally. The hydrogeologic layer representation gives three-dimensional shape to each of the aquifer and nonaquifer layers in the ground-water flow model and provides elevation information on the top and bottom of each layer. Aquifer thicknesses were calculated using the upper and lower surfaces established for these layers. The representation of each hydrogeologic layer was directly entered into the ground-water flow model to define the boundaries and shape of each aquifer layer.

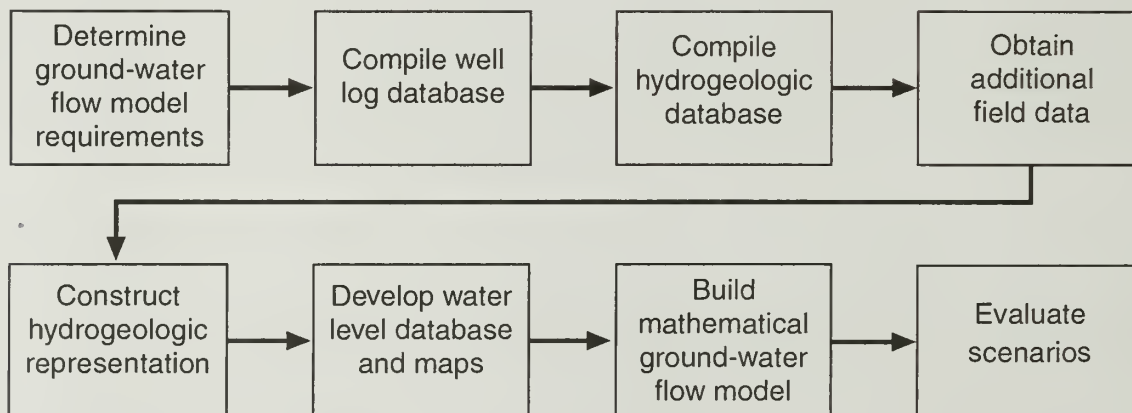


Figure 3. Flow chart showing sequence of activities used to develop the mathematical ground-water flow model

The ground-water-level database included information available from well records containing water-level data (depth to water). Using the known surface elevations, well-screen depths, and calculated well-screen elevations, the water-level data were sorted according to the aquifer layer in which each well was completed. The water level given in the well records was converted to water-level elevation. These elevations were plotted on a map and contoured to provide a potentiometric surface map of each aquifer layer. These maps were the "target" water-level elevations that the ground-water flow model should reproduce when it was calibrated.

Hydraulic properties of the aquifer and aquitard layers that were input into the ground-water flow model included transmissivity, storage coefficient, hydraulic head, and leakage values for each cell in the model. Not all aquifer properties are determinable with a reasonable amount of accuracy from field data. Leakage (recharge) is a good example. Though there are methods available to estimate recharge, it is not directly quantifiable on a cell-by-cell basis for the model. An acceptable approach to determine recharge in a modeling study is through calibration of the model itself, if there is sufficient information available about the ground-water flow system to develop a reliable representation of the aquifer properties.

Maximizing the amount of known data input into a model is the key to increasing confidence in the solution. A model is calibrated by assigning reasonable values for missing data, evaluating the results, and adjusting the assigned values iteratively until the model results match, as closely as possible, the data from field measurements. Details of the methods used to calibrate the ground-water flow model and determine the final outputs are described in the Mathematical Ground-Water Flow Model section of this report.

Quality Assurance

Accurate input data were critical to the development of a valid hydrogeologic representation and a realistic ground-water flow model. Significant steps were taken to ensure quality and completeness of both the information gathered and the final products. For each procedure listed in figure 3, a series of quality assurance steps was followed.

In developing the well log database, water-well information was combined both electronically and in paper form. Because water-well records are maintained by both Surveys, an electronic comparison of the data from each well record was performed. This comparison identified differences in data fields for each well record (i.e., entry errors). From this comparison, a list of water-well records was developed that showed the records which were unique to each Survey (i.e., existed at only one Survey). These data were corrected, as appropriate, for use in the project's databases.

Copies of individual well records also were reviewed to determine if these records contained geologic information specific to the thickness, depth, and lithology of the glacial deposits. Only well records that had such information were included in the hydrogeologic database. The accuracy of the well locations and elevations reported in the database then were checked. For well records without complete geographic coordinates, location information was gathered by matching the well owner's name given on the well record with that in the plat book for the well location. If the land-surface elevation was missing from the record, it was estimated by locating the well site on the appropriate 7.5-minute topographic map. If topographic relief was great, or if the well site or owner were not identifiable, the record was not used. In the few cases when the geologic information from a well record was considered to be particularly important, the driller was contacted to help identify the well's location.

The geologic information on each log was evaluated, and preliminary maps of the hydrogeologic layers were produced. The maps underwent a rigorous review by project staff and external reviewers. The review process was greatly enhanced by the three-dimensional geologic model software EarthVision, which allowed reviewers to view any cross sections or layers and to identify poorly interpreted areas more easily. The review

process for the data elements in this study involved staff from both Surveys and evolved into a successful system of checks and balances. The various areas of expertise of the reviewers proved to be an integral part of developing a consistent, coherent hydrogeologic interpretation.

In the hydrogeologic characterization report developed for this study, Herzog et al. (1995a) described the geology of the study area as a two-aquifer-layer system. The water-level data were grouped into these two aquifer layers for that report. Reinterpretation into three aquifer layers was completed after an initial modeling attempt revealed that the combining of layers into two units provided an inadequate representation in the model. The model created artificially high and low water levels that made the model unstable. In the revised three-aquifer-layer conceptual model developed in this report, the data were re-evaluated to place the water-level information with the appropriate aquifer layer. The well screen elevations for each well were entered into EarthVision to determine the aquifer layer to assign each water-level measurement. In some cases, well screens were found to be in a nonaquifer layer. These logs were examined in more detail to evaluate why this occurred, because the well screen would not actually be in a nonaquifer material. In most cases, the problem revolved around the depth of the hole drilled being different than the actual depth of the screen. In other words, in some cases the driller placed the well screen above the bottom of the well boring. In practice, well drillers generally put the screen at the bottom of the boring. The rest could not be resolved, however, because they were an artifact of grouping the shallow sand-and-gravel deposits into two layers. These were resolved by assigning the water level to the closest aquifer layer. Taking this additional step provided a check of the hydrogeologic mapping, strengthening the final interpretation, in addition to providing a method to evaluate the water-level data properly.

Details of the measurement of the water levels from private wells and for the aquifer pumping tests are described in Herzog et al. (1995a). Water-level data taken from drillers' logs often are more suspect because the exact measurement method, accuracy of the reading, and well status are not recorded on the well logs. Water levels from drillers' logs were used only when there were several wells in close proximity that corroborated a single value, or if no other data were available. Historical aquifer test data in the Aquifer Properties Database at the ISWS were reviewed to ensure that the analysis followed established procedures and to confirm the results of the original analysis.

The ground-water flow model required values for water level and hydraulic properties at every cell in the model. A cell is a small portion of the model at which the equations of ground-water flow are solved when the model is run. The model developed for this project has 122 rows of cells and 146 columns of cells for each layer. Because the model has more than 100,000 cells, it was not possible to provide unique values for each. Hydraulic properties were determined from the two aquifer tests conducted during the aquifer characterization portion of this study and from the 20-30 aquifer test results from community well tests in the Aquifer Properties Database at the ISWS. Although these data were valuable, variations in hydrogeology found during test drilling for the aquifer characterization study suggest that the hydraulic properties could vary significantly throughout the study area. In the absence of data similar to that obtained from the two aquifer tests, the hydraulic properties of the earth materials in the rest of the study area were estimated based on the physical characteristics of the earth material. These limitations and other aspects of quality assurance for the ground-water flow model are described further in the Mathematical Ground-Water Flow Model section of this report.

HYDROGEOLOGY

Because sand-and-gravel aquifers within the glacial deposits have great potential for yielding a large ground-water supply in the model area, the detailed characterization of the sequence of glacial deposits is critical for this study. An understanding of the character and distribution of the glacial deposits is crucial to understanding the ground-water flow system. Understanding the topography of the bedrock surface is just as

important because the topography influences the location and areal extent of the major aquifer and, as a hydraulic boundary, affects the ground-water flow system.

Geologic Database

Well Records

The hydrogeology of the 1,100-square-mile model area was described using information gathered from well records on file at the ISGS and ISWS. These records include logs of water wells, engineering boreholes, and test borings for coal, oil, and gas. Although more than 8,000 well records were available for the model area, only 2,278 of them, whose logs provided data on the thickness, depth, and lithology of the glacial deposits, were used in the hydrogeologic database. The logs for the majority of well records were not helpful for the geologic interpretations needed for this study because they were for very shallow test holes, or they provided little information about the nature of the glacial deposits. The database included the logs of the 26 test holes drilled for aquifer characterization phase of this study (Herzog et al., 1995a) and the one additional well drilled during the modeling phase of the study.

The most accurate and detailed geologic descriptions came from the geologists' logs for the 27 test holes drilled for the aquifer characterization study (Herzog et al., 1995b), and from logs of the test wells drilled for municipal water supplies (e.g., tests for the Normal well field). The locations and land-surface elevations of these well sites have been surveyed and are highly accurate. Logs from test drilling provided detailed geologic descriptions of all deposits encountered. The test drilling usually terminated in bedrock. Downhole geophysical logs (Herzog et al., 1995b) confirmed the type and sequence of earth materials. Additionally, many of the test-hole samples have been analyzed for grain-size distribution and clay mineralogy and studied under a microscope to provide detailed sample descriptions. Therefore, the data from these logs provide the key control for establishing the regional stratigraphic framework. The high-quality logs of test holes and test wells provided the basis for interpreting the information from the less complete logs in the water-well records. The usually brief descriptions of geologic materials given in drillers' logs became more useful when compared to the descriptions made by geologists in the logs of key stratigraphic-control test holes. The drillers' descriptions were then put into stratigraphic context and used to improve geologic understanding of the area.

The information from the 2,278 water-well records used in this study varied widely in quality of location information and geologic data provided in the log. Many of the well records contained no land-surface elevation for the location, so elevations for these well locations were estimated from 7.5-minute topographic quadrangles with an estimated accuracy of within 5 feet. The brief descriptions for various geologic materials typically used in the logs, such as "sand," "clay," and "soil," gave limited information from which to identify stratigraphic units.

Because the Sankoty-Mahomet Sand aquifer can meet most water-supply needs from the upper part of the aquifer, fewer than 500 water wells penetrated the full thickness of this aquifer. However, many drillers' logs recorded every change in earth material encountered, including sand seams as thin as several inches, soil, peat, and other organic material. Buried soils, which formed between episodes of glaciation, were extremely useful in correlating geologic units between well locations. Because the water-well drillers' primary goal is to find a water supply, their logs tend to emphasize sand-and-gravel deposits that are capable of supplying a household with water. This tendency is increased because the rotary drilling method used by most water-well drillers does not allow identification of thin sand seams. Thus, thin sand seams that are thought to be common in the model area may not be noted on the logs. Also, water-well contractors who plan to drill into the widespread and reliable water source of the Sankoty-Mahomet Sand aquifer may not describe the shallower sands. Therefore, the lack of sand as recorded in a driller's log may mean either no sand was encountered or none was described.

Logs also were available in the records of test holes drilled for coal or oil and gas. Because these test holes targeted horizons deep within the bedrock, the logs typically did not provide useful descriptions of the shallower glacial deposits. In most cases, however, they had accurate information on the depth to bedrock from which the elevation of the bedrock surface was determined. Because the bedrock surface is the lowest possible boundary for the Sankoty-Mahomet Sand aquifer, these data supplemented the data available on the aquifer's thickness. Location information for these test holes is typically very accurate.

On copies of all 2,278 well records, descriptions and depths of "marker" beds (e.g., wood, peat, "organic," green clay, etc.) and intervals of coarse-grained sediment (sand, sand and gravel, or gravel) were highlighted in different colors to facilitate entry into a spreadsheet. Unique well identification numbers, locations by legal description, land-surface elevations, and total depths from the "header" database were entered into the spreadsheet. Columns were added for the depths, thicknesses, and elevations of the three aquifer layers, depths and elevation of "marker" beds, and depth and elevation of the bedrock surface. The process of assigning the coarse-grained units to one of three aquifer layers is described in the section on Hydrogeologic Layer Development. A portion of the completed database that exemplifies the differences in data quality is in appendix B.

Although the logs in 2,278 well records were used, most did not provide comprehensive information on the depth and thickness of every unit. Table 1 shows the number of values entered into the spreadsheet for each aquifer layer and "marker" bed. These include values from logs, inferred values, and constraining values. All three types of values were necessary because the ground-water flow model characterized all layers as continuous over the entire model area. Inferred values of thickness were predominately values of zero thickness for a particular sand unit from logs with sufficient detail to determine with confidence that the sand unit was absent at that location. Use of these inferred zero values is the reason that table 1 shows considerably more values of thickness of the Glasford and lower Glasford/upper Banner aquifer layers than for the elevations of the tops of these layers. No values were estimated for the tops of the aquifer layers if they were interpreted to be absent. When the Sankoty-Mahomet Sand aquifer was known to be absent, the elevation of the top of this aquifer was inferred to be less than 1 foot above the elevation of the bedrock surface to provide the needed continuity for this aquifer layer in the ground-water flow model.

Constraining values were thicknesses derived from the logs of wells that did not fully penetrate the thickness of an aquifer layer. These values were used in mapping the hydrogeologic layers to ensure that the

Table 1. Number of Values for Depth and Thickness for Each of the Modeled Hydrogeologic Units

<i>Hydrogeologic unit</i>	<i>Number of values</i>
First marker: top of the Glasford Formation (Sangamon Soil, Robein Silt, etc.)	674
Elevation of the top of the Glasford aquifer layer	978
Thickness of the Glasford aquifer layer	1,536
Second marker: top of the Banner Formation (Yarmouth Soil, Lierle Clay, etc.)	279
Elevation of the top of the lower Glasford/upper Banner aquifer layer	1,101
Thickness of the lower Glasford/upper Banner aquifer layer	1,331
Elevation of the top of the Sankoty-Mahomet Sand aquifer layer	1,221
Thickness of the Sankoty-Mahomet Sand aquifer layer	498
Elevation of the bedrock surface	371

aquifer layer was given at least that measured thickness. For example, if a driller's log showed at least 150 feet of sand and, if not using this value would cause the thickness of the sand to be shown as less than 150 feet at that location on the thickness map, the value was included in the database as the thickness of that sand. Because these constraining values were used in conjunction with measured thickness values, they increased the map's accuracy. If the thickness map indicated that the sand thickness was greater than 150 feet thick at that location, the point was not used.

New Field Data

A test hole was drilled for the model study to supplement the test drilling completed for the aquifer characterization study (Herzog et al., 1995b). Test hole MTH-27 was drilled into bedrock at a site located in the northwest corner of the NE¼ Section 36, T23N, R2W. The specific purpose for drilling at this location was to determine the thickness of the Sankoty-Mahomet Sand aquifer east of test hole MTH-12, located about 1¼ miles west of MTH-27. Herzog et al. (1995b) found the Sankoty-Mahomet Sand aquifer at MTH-12 to be only 12 feet thick. Because MTH-12 is located in the middle of the study area where the bedrock surface is deepest, it was important to determine how far this limited thickness of aquifer extended to the east. Test hole MTH-27 was drilled to a depth of 295 feet. Bedrock was encountered at a depth of 280 feet. The field methods described in Herzog et al. (1995a) were used to collect and describe the samples in the field. The Sankoty-Mahomet Sand aquifer was found to be 102 feet thick, which fit the thickness of between 100 and 125 feet predicted by the aquifer characterization study (Herzog et al., 1995a). An observation well was installed in the Sankoty-Mahomet Sand aquifer in this test hole, properly developed, and added to the network for routine water-level measurements. The geologists' and drillers' logs are presented in appendix C.

Physical Features of the Landscape

Most of the model area is located in the Bloomington Ridged Plain-Till Plains Section-Central Lowland Province (Leighton et al., 1948), where deposits from the last continental ice sheet form the land surface. Erosion and deposition have defined the major features of present-day landforms, which subsequent weathering and fluvial erosional processes have altered to their present shape. The broad, arcuate ridges of glacial end moraines, which mark former ice-marginal positions, trend northwest to southeast across the area. The southwestern quarter of the model area is beyond the moraine that marks the limit of the Wisconsin Episode glaciation. This area is in the Springfield Plain-Till Plains Section-Central Lowland Province (Leighton et al., 1948), where deposits of an older glacial episode form a distinctively flat land surface.

Modern streams in the model area mostly have a northeast to southwest orientation and comprise two major drainage systems. The major stream in the northern half of the model area—the Mackinaw River—flows southwest between Goodfield and Congerville, past Mackinaw toward Hopedale where it turns west to exit the model area (figure 1). Sugar Creek and Kickapoo Creek comprise the major drainage system in the southern half of the model area (figure 1). Both drainage systems contain numerous minor tributaries.

Bedrock Geology

Shale and relatively thin layers of sandstone, limestone, and coal of the Carbondale Formation (Pennsylvanian Period) comprise the upper part of the bedrock in the model area. No significant aquifers are found within the bedrock. Because mineral content of the ground water in the bedrock increases with depth, water 50 to 100 feet below the bedrock surface may be too highly mineralized for most uses. Bedrock units within the study area cannot produce enough water for a municipal supply, although the relatively fine-grained sandstone or fractures in the limestone and coal may yield sufficient quantities for domestic use. The shale does not typically yield enough water for a domestic supply.

The bedrock surface is a major hydrologic constraint within the model area and forms the base on which the ground-water flow system in the glacial deposits rests. The areal extent of the Sankoty-Mahomet Sand aquifer, the principal glacial-drift aquifer in the area, is influenced by the topography of this surface (figure 4). The confluence area of the buried Mackinaw and Mahomet Bedrock Valleys underlies much of the model area. Margins of the bedrock valleys have been approximated by the 500-foot bedrock elevation contour (Kempton et al., 1991) because this elevation typically marks the top of the Sankoty-Mahomet Sand Member. Bedrock elevation locally exceeds 550 feet on bedrock uplands adjacent to the two bedrock valleys. The shape of the bedrock surface in the confluence area was interpreted as a broad, relatively flat-bottomed basin (for example, see figure 14 in Kempton and Visocky, 1992). However, recent studies in the area that involved significant test drilling (Wilson et al., 1994; Herzog et al., 1995a and b), as well as other recent mapping projects in McLean and Champaign Counties (Kempton et al., 1995), have greatly improved topographic interpretation of the bedrock surface within the confluence area. The configuration of the bedrock surface on uplands adjacent to the bedrock valleys was interpreted from drillers' logs included in water-well records.

The bedrock surface in the model area also includes two other valleys: the northeast-southwest trending Danvers Bedrock Valley in the northeastern corner of the model area and the northwest-southeast trending Kenney Bedrock Valley along the south edge of the model area. The model area was selected to encompass bedrock valley walls and uplands, where the bedrock surface elevation commonly exceeds 500 feet. Several bedrock hills occur within the central part of the model area, the highest of which is the hill south of Hopedale with an elevation above 500 feet. These hills are significant because they interrupt the thickness and continuity of the Sankoty-Mahomet Sand aquifer.

Glacial Geology

This section provides a description of the glacial geology of the model area, using the stratigraphic terminology of Hansel and Johnson (1996). An understanding of glacial geology is essential for simulating the characteristics and continuity of the aquifer system. The approach used for interpretation of the glacial geology input into the ground-water flow model is described in the section on Hydrogeologic Layer Development.

Continental glaciers that originated in north-central Canada modified the topography of the preglacial bedrock surface of Illinois by initially deepening the bedrock valleys through erosion and eventually filling them with glacial and proglacial sediment. At least three cycles of advances and retreats of the ice sheets left behind layers of glacial sediment. The older ice sheets covered more of the state than the more recent ice sheets. Sand-and-gravel outwash deposited by meltwater rivers flowing from the pre-Illinois Episode glaciers filled most of the deeper parts of the bedrock valleys. Each glacial advance and retreat modified the existing landscape through erosion and by deposition of sediment directly from glacial ice, meltwater streams, and proglacial lakes. Shifting ice margins modified drainage patterns, and lakes formed where ice or glacial sediments blocked the drainage. Glacial events ceased to affect Illinois directly about 12,000 years ago and left more than 400 feet of glacial and proglacial deposits in some parts of the model area. The glacial sediments are thinnest above the bedrock uplands that form the sides of the bedrock valleys and over bedrock highs within the valleys. They are thickest within the deepest sections of the bedrock valleys. The proglacial deposits include a thick layer of sand and gravel directly overlying bedrock in most of the deeper parts of the Mahomet and Mackinaw Bedrock Valleys and their tributaries. These units are called the Mahomet and Sankoty Sand Members of the Banner Formation.

The glacial deposits include diamicton, outwash, and lacustrine sediments. Diamicton (commonly referred to as till) is unsorted, nonstratified sediment with a wide range of particle sizes (e.g., clay to pebbles, cobbles, and boulders) deposited adjacent to or directly from ice. Outwash consists mainly of bedded sand and gravel deposited in proglacial streams and rivers. Outwash may be confined between valley walls in long, narrow deposits called valley trains, or it may be spread over a large area as a flat or gently sloping sheet of sediment

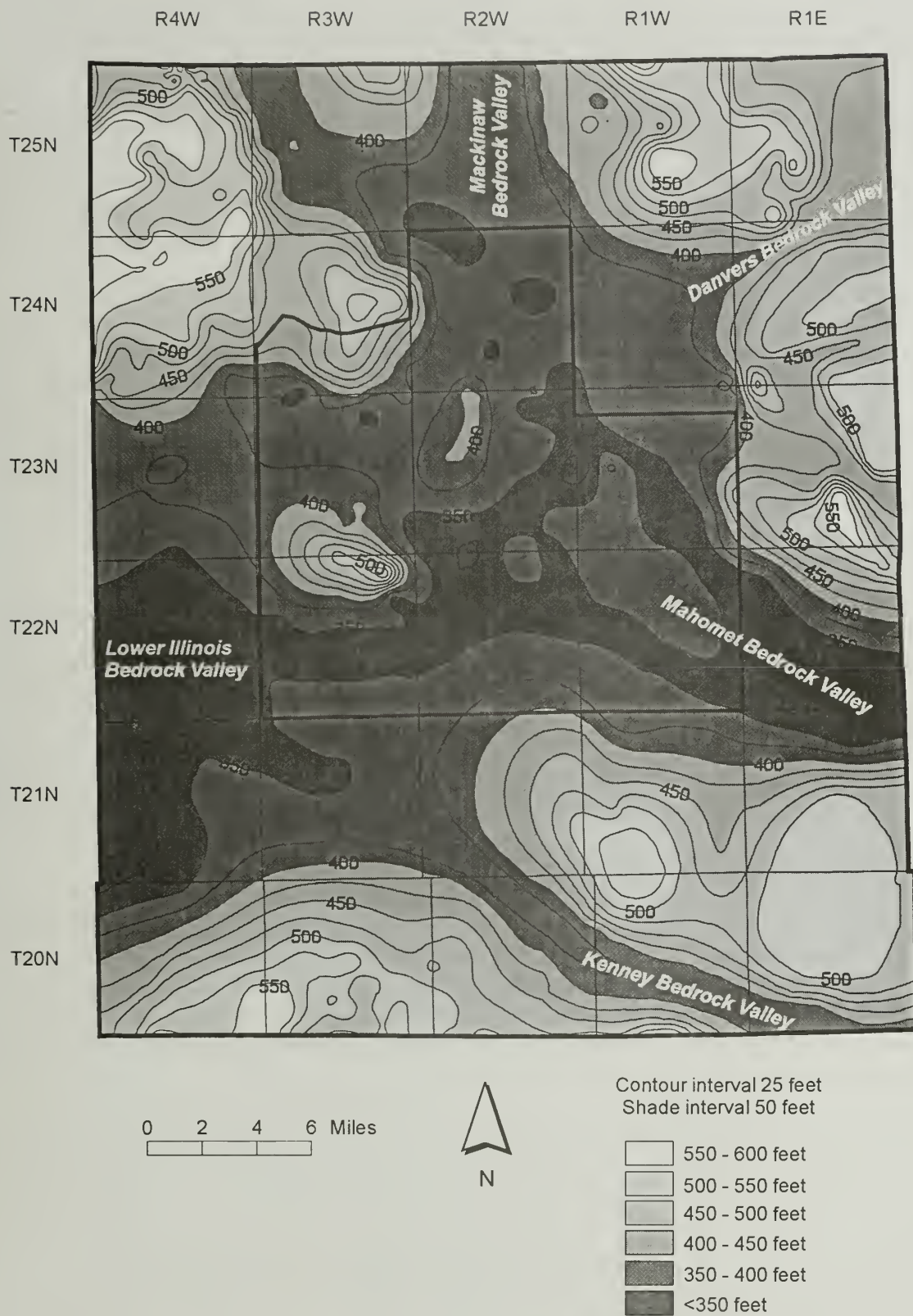


Figure 4. Elevation of the bedrock surface

called an outwash plain. Lacustrine sediments are fine-grained clays and silts deposited in the relatively quiet water of lakes.

Glacial and related deposits are identified, distinguished, and classified based on their color, lithology, stratigraphic position, and age. Buried weathered zones (soils), some containing organic-rich horizons, serve as important marker beds. Buried soils indicate periods of warmth and weathering between glaciations, and mark significant discontinuities (unconformities) in the sedimentary record. They are used to separate the glacial deposits into distinct stratigraphic units.

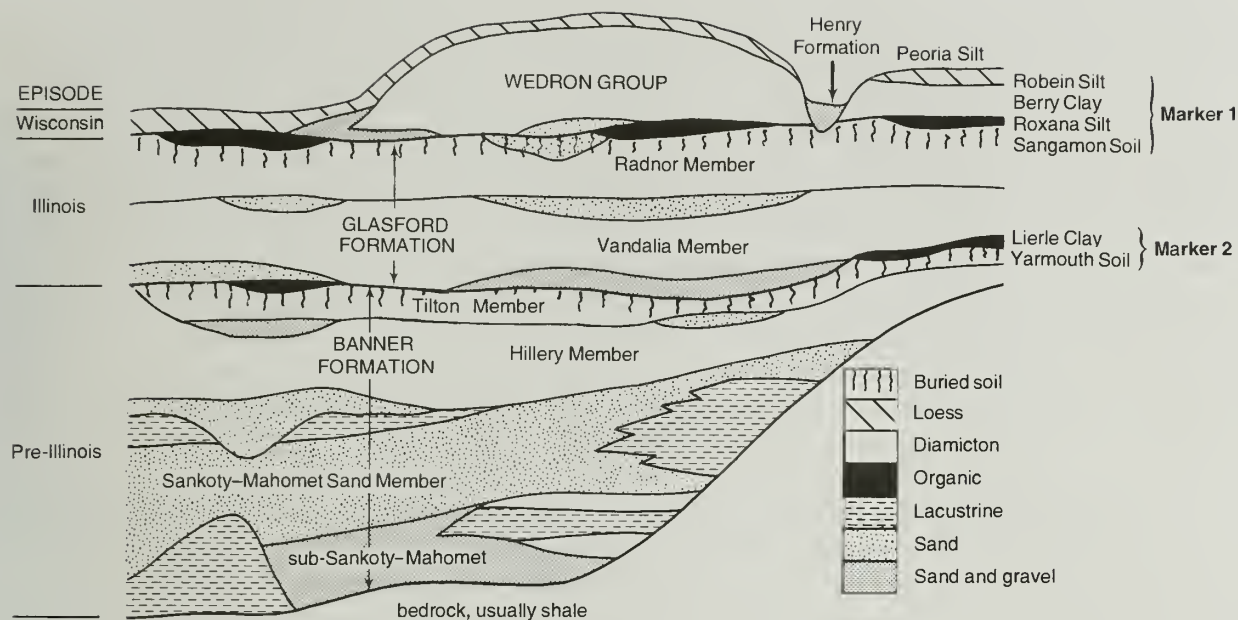
The glacial deposits are grouped into three major stratigraphic units (figure 5a). From oldest to youngest, these are the Banner Formation, the Glasford Formation, and the Wedron Group. These units are separated locally by well developed soils and organic horizons. The Yarmouth Soil and Lierle Clay separate the Banner Formation from the overlying Glasford Formation. The Sangamon Soil, Berry Clay, and Robein Silt separate the Glasford Formation from the overlying Wedron Group. Each of these major stratigraphic units contains aquifers composed of sand-and-gravel deposits, as well as aquitards composed of diamictos and lake sediments. As shown in figure 5b, the stratigraphy of the aquifers was simplified into three aquifer layers as discussed in more detail in the section on Hydrogeologic Layer Development.

Banner Formation

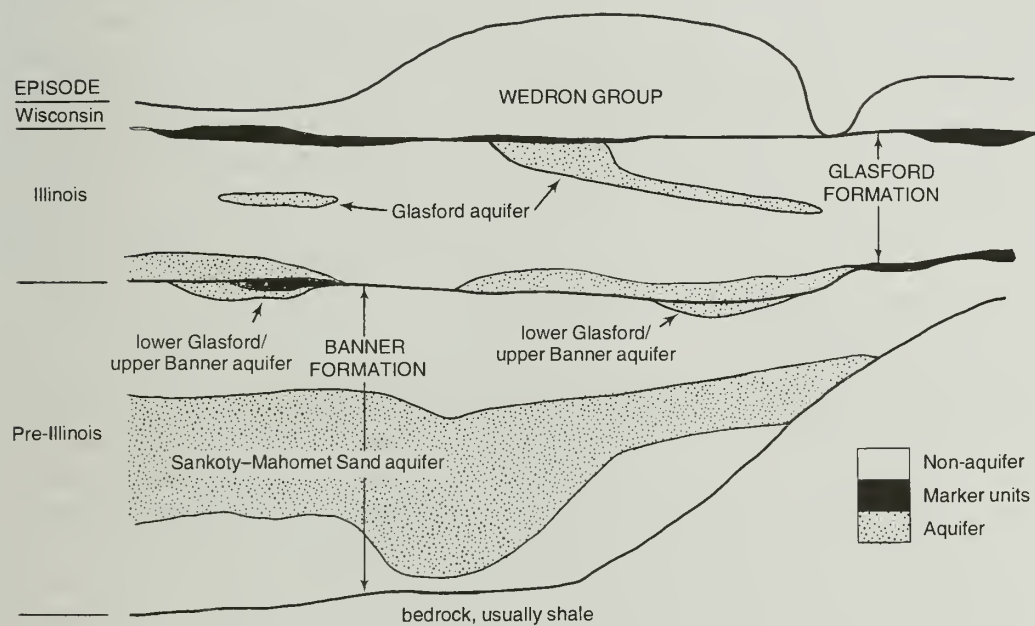
The Banner Formation, thought to have been deposited more than 500,000 years ago, is the lowermost glacial unit in the model area. It rests on the bedrock surface. The top of the Banner Formation is often marked by a discontinuous, buried weathered zone (Lierle Clay in figure 5a). Otherwise, the Banner Formation directly underlies the younger Glasford Formation. The top of the Banner Formation ranges in elevation from approximately 650 feet in the northeastern part of the model area to less than 500 feet in the southeastern corner and the west side of the area. It is generally draped over the bedrock surface and is deepest where the bedrock is deepest. The Banner Formation is generally absent where the bedrock surface is high; it tends to be thickest along the axes of the Mackinaw and Mahomet Bedrock Valleys. In a few localized areas, it may be more than 175 feet thick.

At the base of the Banner Formation lies the major sand-and-gravel unit in the model area. In the Mackinaw Bedrock Valley and extending to the Peoria area, this unit was named the Sankoty Sand Member. In the Mahomet Bedrock Valley, it was named the Mahomet Sand Member (Willman et al., 1975). However, it is referred to as the Sankoty-Mahomet Sand Member for this report (figure 5a) for two reasons: (1) the model area includes the confluence area of the Mackinaw and Mahomet Bedrock Valleys where these two members are mostly indistinguishable, and (2) the Mahomet and Sankoty Sand Members form a single hydrologic unit, the Sankoty-Mahomet Sand aquifer (Visocky and Schicht, 1969).

Before the aquifer characterization study, the lower part of the Banner Formation was thought to be composed of a single sand-and-gravel deposit, the Sankoty-Mahomet Sand Member. Based on geologic data, Herzog et al. (1995a and b) informally subdivided this unit into the lower, sub-Sankoty-Mahomet sediments and the upper, Sankoty-Mahomet Sand Member (figure 5a). Diamictos or fine-grained lacustrine sediments locally separate the two sand-and-gravel deposits. If the diamictos or lacustrine sediments are absent, it is difficult to distinguish the sand and gravel of the sub-Sankoty-Mahomet sediments from that of the directly overlying Sankoty-Mahomet Sand Member, although the lower unit is generally coarser grained. In some areas in southwestern McLean County, and near the center of the model area, fine-grained lacustrine sediments are found beneath the sand and gravel of the sub-Sankoty-Mahomet sediments (Herzog et al., 1995a).



a. Sequence of geologic material in the study area



b. Simplification used in hydrogeologic representation

Figure 5

The fine-grained facies of the sub-Sankoty-Mahomet sediments directly underlying the Sankoty-Mahomet Sand Member are typically thin or absent over much of the model area. Consequently, the sand and gravel of the sub-Sankoty-Mahomet sediments are hydraulically connected to the Sankoty-Mahomet Sand Member and form the Sankoty-Mahomet Sand aquifer. The thickness of the aquifer generally follows the trend of the bedrock-surface topography. The aquifer is more than 50 feet thick throughout much of the model area and more than 150 feet thick in the deeper parts of the bedrock valleys, such as between Tremont and Hopedale, southeast of Hopedale, and east of Mackinaw. It thins near the valley edges and over several of the bedrock highs in the center of the study area; it is completely absent at the bedrock high south of Hopedale. The aquifer typically is thin in areas in which the elevation of its top is below 425 feet (Herzog et al., 1995a). In these areas, such as southwestern McLean County and the center of the model area, fine-grained sediments are commonly found instead of the Sankoty-Mahomet Sand aquifer. The occurrence of lacustrine sediments within the broad bedrock valley suggests that lakes occupied parts of the bedrock valleys during the pre-Illinois Episode (Herzog et al., 1995a), possibly because ice blocked the main channels.

Two diamictons, the Hillery and Tilton Members, comprise the upper part of the Banner Formation, above the Sankoty-Mahomet Sand aquifer. Two locally significant sand-and-gravel deposits occur, one separating these two diamictons and one at the top of the Banner Formation or the bottom of the Glasford Formation. These aquifers are generally less than 25 feet thick, but they are absent in some places and more than 50 feet thick at others. They are widespread and locally are a significant source of water; they supply water to many small communities and domestic wells in the model area.

Glasford Formation

The Glasford Formation, which is thought to have been deposited between 180,000 and 125,000 years ago during the Illinois Episode, overlies the Banner Formation and underlies the Wedron Group. The top of the Glasford Formation was inferred by the presence of the Robein Silt, an organic-rich, distinctive marker bed found above the Glasford Formation over much of the state. Where the Robein Silt is absent, the sediments deposited during the Sangamon Interglacial Episode that followed the Illinois Episode are commonly found in Glasford Formation deposits. The Robein Silt is distinctive enough to have been noted in many of the well logs and sample sets from throughout the area (Herzog et al., 1995b). The elevation of the top of the Glasford Formation ranges from above 700 feet on the eastern side of the model area to lower than 550 feet on the west. Total thickness of the Glasford Formation ranges from less than 25 feet to more than 125 feet; it is generally thickest where the elevation of its top is highest.

Two diamicton units (the Vandalia Member and the overlying Radnor Member) comprise most of the Glasford Formation. However, outwash deposits of sand and gravel locally are thick enough to be significant aquifers. These deposits generally are found at the base of the Vandalia Member and between the Vandalia and Radnor Members (figure 5a). The total thickness of the sand-and-gravel deposits above the Sankoty-Mahomet Sand aquifer in both the Banner and Glasford Formations within the confluence area is generally less than 25 feet, an inadequate thickness for high-capacity wells. In some parts of the model area, however, basal Vandalia sand and gravel directly overlie the Sankoty-Mahomet Sand aquifer and form one thick aquifer. The most notable example is found in the western side of the model area north of the Mackinaw River, where a sand-and-gravel thickness of 199 feet was found in test hole MTH-7 (Herzog et al., 1995b).

Sand-and-gravel deposits locally separate the Glasford Formation (figure 5a) into two parts. Some narrow channels containing sand-and-gravel deposits occur at the top of the Radnor Member. Sand-and-gravel deposits at the base of the Radnor Member are more widespread. Sand-and-gravel deposits within the Glasford Formation are widely distributed across the model area. Typically they are thin and rarely exceed 25 feet in thickness. When present, however, they may be capable of furnishing a domestic water supply.

Wedron Group

The Wedron Group (figure 5a) represents Wisconsin-age deposits (deposited between 25,000-12,000 years ago in this area) and is the principal surficial unit throughout most of the model area. It directly overlies the Glasford Formation, the Robein Silt, or the Roxana Silt. The Wedron Group is composed mostly of diamicton; but it contains sand-and-gravel deposits that are typically thin, discontinuous, and very limited in areal extent. The Wedron Group is absent in the western and southwestern part of the model area, which is beyond the limit of the Wisconsin Episode glaciation. It is greater than 100 feet thick in the Bloomington and LeRoy Moraines, which appear as arcuate ridges across the landscape; and it is thin or absent where the Mackinaw River and Sugar Creek have eroded it.

The Tiskilwa Formation, a pinkish gray, pebbly clay till, is the lowermost diamicton of the Wedron Group. The Tiskilwa Formation extends from the land surface downward to an elevation of about 600 feet. A thin, younger Wedron diamicton unit, the Lemont Formation, locally overlies the Tiskilwa Formation. Thin lenses of sand and gravel are locally present between the two diamictons. Because of their irregular shapes, these thin, discontinuous deposits are difficult to map. They offer little or no potential for the development of a municipal water supply, but they may have limited potential as the source of supply for large-diameter domestic wells. Although no attempt was made to map these shallow sands, their depths and elevations were used in the hydrogeologic database to track specific locations at which ground-water recharge may be enhanced. A thin layer of sand and gravel is found at the base of the Wedron Group in a few places. It was included in the Glasford aquifer layer for localities in which it directly overlies the Glasford Formation.

The Peoria Silt is a late glacial wind-blown silt (loess) that covers the upper surface of the Wedron Group over much of the model area to a depth of as much as 10 feet (Herzog et al., 1995a). Surficial deposits of sand and gravel of the Henry Formation overlie the Wedron Group. These outwash deposits are found along the principal streams in the model area and along some end moraines that form the outer margin of the Wedron Group. Although the sand-and-gravel deposits of the Henry Formation are typically thin and limited in areal extent, they may be as thick as 60 feet (Herzog et al., 1995a).

Hydrogeologic Layer Development

A necessary step in designing the mathematical ground-water flow model was to establish a series of layers representing aquifers and aquitards. This step is difficult when working with a glaciated terrain because glacial deposits are heterogeneous and usually are not deposited as widespread, uniform layers. Aquifer material is deposited by glacial meltwater, which commonly followed stream valleys. This depositional environment tends to make them linear features. The hydrogeologic model developed for this project is a simplified three-dimensional representation of the subsurface earth materials from the land surface to the bedrock surface. The bedrock acts more or less as an aquitard and, thus, is the deepest geologic unit depicted in the ground-water flow model. The hydrogeologic layer representation is based on geologic data from well logs and maps that show the top, thickness, and bottom of each sedimentary unit (figure 6). This three-dimensional depiction allowed geologists to view their interpretations of the relationships among the mapped layers from many directions to assure that the interpretations were internally consistent (i.e., to assure that the older units were always plotted below younger units and that they did not intersect unless a real feature, such as a bedrock hill, was identified from the data).

Digital files representing the thickness or elevation of each unit were input into computer mapping and geologic representation programs. All of the geologic maps for the model area were developed electronically using the GIS software package ARC/INFO and the three-dimensional representation software EarthVision.

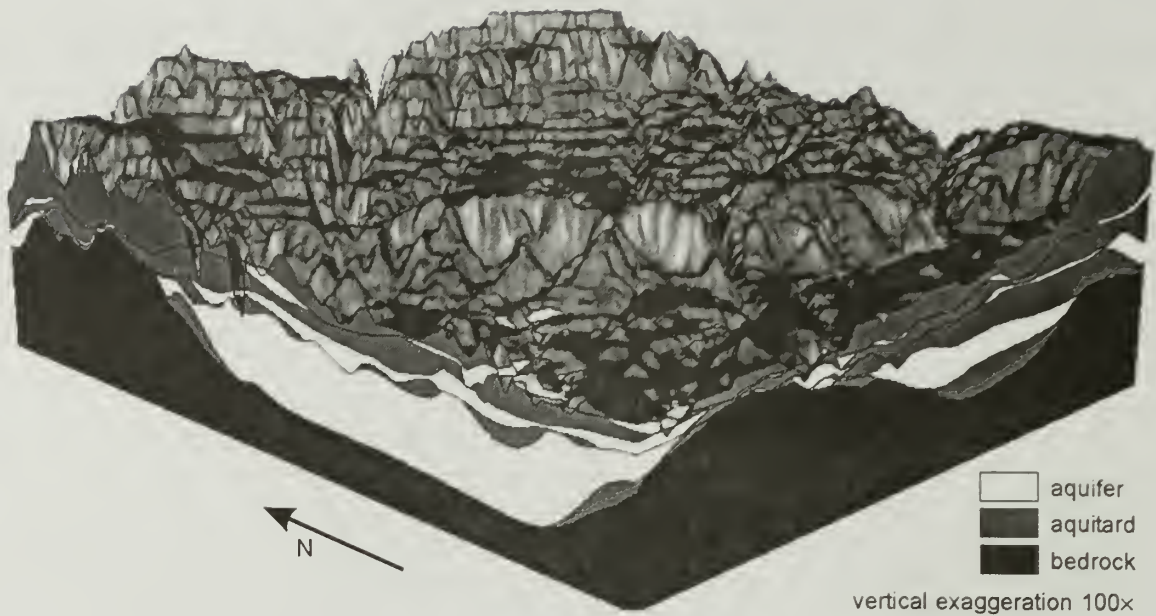


Figure 6. Three-dimensional hydrogeologic representation of the model area viewed from the southwest corner

Generation of a hydrogeologic representation, such as the one shown in figure 6, is an iterative process involving several steps (figure 7).

Two-dimensional surfaces of the tops of geologic units were created using data from the well records. Each two-dimensional surface is represented as a grid, which is a quarter-mile-spaced rectangular array of points that contains elevation values. A minimum tension gridding algorithm was used to represent the values of the input data as closely as possible to create a two-dimensional surface with as little curvature as possible.

Maps of two "marker" beds were produced. "Marker" bed 1 is the Sangamon Soil-Robein Silt and marks the contact between deposits of the Illinois and Wisconsin Episodes of glaciation (figure 8). "Marker" bed 2 is the Yarmouth Soil-Lierle Clay and marks the contact between the deposits of the pre-Illinois and Illinois Episodes of glaciation (figure 9). The "marker" beds depict the approximate land-surface topography (now buried) that existed during warm periods between two major episodes of glaciation. They were developed mainly to provide a geologic basis for assigning the sand-and-gravel deposits shown on well logs to one of three aquifer layers in the hydrogeologic model. The top elevations of the upper two of these aquifer layers, the Glasford and lower Glasford/upper Banner aquifers, are near the elevation of the "marker" beds, providing a rationale for assigning sand-and-gravel deposits to these aquifer layers. The top elevation of the lowest aquifer layer (Sankoty-Mahomet Sand) is noticeably lower than either of the "marker" bed elevations, and its thickness frequently extends to bedrock.

Because these maps are interpretations not required of the project, they were not extensively reviewed. The "marker" bed maps, however, provided insight about the shape of ancient land surfaces. For example, the map of "marker" bed 2 (figure 9) suggests that a stream flowed from east to west through the model area, bisecting the ancient land surface beneath what is now the Normal well field.

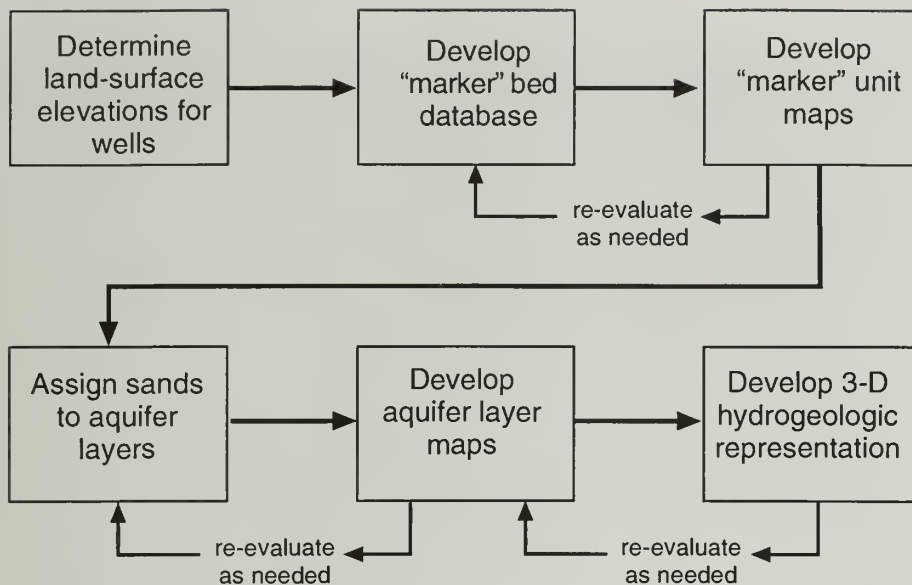


Figure 7. Steps in developing hydrogeologic layer representation

Maps were made of the elevation of the tops and bottoms of the three aquifer layers and the elevations of land and bedrock surfaces. These maps were reviewed and revised several times, checking for anomalous data points. In areas of apparent elevation anomalies, the original well logs were checked for accurate entry of the data into the spreadsheet, accurate land-surface elevation, and accurate well location. In some cases, coarse-grained units were reassigned to different aquifer layers.

The grids of the individual surfaces were combined into a three-dimensional hydrogeologic representation. The software allowed surfaces to be defined as "depositional," in which one layer is draped over another, or as "erosional," in which one layer is eroded into a lower surface. In the model area, only the land surface layer was set as erosional; all the other layers were depositional. After the three-dimensional hydrogeologic representation was assembled, it was possible to view the surface of the layers and an unlimited number of cross sections through the model area. Dozens of cross sections of the model were reviewed for internal consistency. When inconsistencies were discovered, data were rechecked using the same procedure that was used for checking anomalous points on the maps of aquifer layers. This required revision and review of the maps, and re-running of the three-dimensional hydrogeologic representation. Twelve revisions of each aquifer-layer map and five iterations of the three-dimensional compilation were required before the project scientists and their reviewers were satisfied that the hydrogeologic layers were a reasonable representation of the subsurface hydrogeology to be used as suitable input for the ground-water flow model. Acceptance of the compilation was based on its agreement with input data, and reasonableness of the appearance of the surfaces and various cross sections.

Hydrogeologic Layers

For the ground-water flow model, the complex glacial geology had to be simplified into a sequence of three aquifer and three nonaquifer layers. The geologic unit below the Sankoty-Mahomet Sand aquifer, fine-grained material or bedrock, was a fourth nonaquifer. The number of aquifer layers was restricted because water-level data were required for each aquifer layer so the model could be calibrated to those data values. To make the



Figure 8. Elevation of "Marker" bed 1

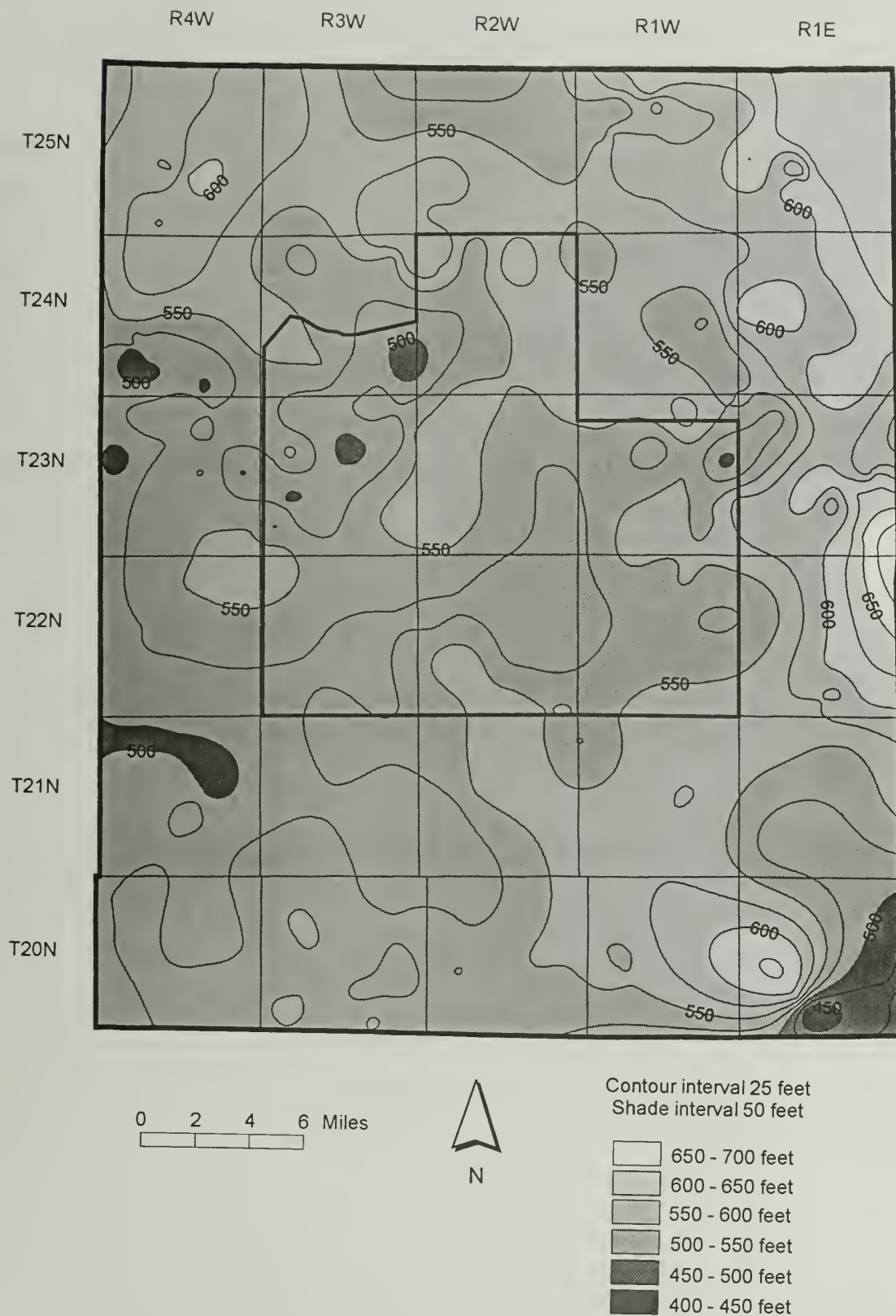


Figure 9. Elevation of "Marker" bed 2

model as representative of actual conditions as possible, three aquifer layers were used (figure 5b). These layers were viewed both as a three-dimensional diagram (figure 6) or as maps of the tops and thicknesses of each layer (figures 10-15).

The Sankoty-Mahomet Sand aquifer, the main aquifer in the model area, also is the deepest aquifer layer in the model area. This layer directly overlies the bedrock surface or lacustrine material. For the hydrogeologic model, the aquifer was mapped as a continuous unit with any interbedded fine-grained material assigned to the underlying aquitard layer to preserve the thickness of the aquifer layer. The elevation of the top of this aquifer layer ranges from more than 525 feet over some of the bedrock highs to below 400 feet in reaches of the bedrock valleys where it is extremely thin (figure 10). The calculated aquifer-layer thickness (figure 11) is more than 50 feet throughout much of the model area and is more than 150 feet in the deeper parts of the bedrock valleys, such as between Tremont and Hopedale, southeast of Hopedale, and east of Mackinaw. As would be expected of a valley-filling fluvial deposit, the aquifer layer thins near the valley sides and over several of the bedrock highs in the center of the study area. The bedrock hill south of Hopedale intrudes completely through the aquifer.

Sand-and-gravel deposits in the Banner Formation above the Sankoty-Mahomet Sand aquifer and at the base of the Glasford Formation were combined to form the lower Glasford/upper Banner aquifer layer. These deposits are close together vertically, appear to be hydraulically connected based on water-level data, and are difficult to distinguish. This aquifer layer straddles the contact between deposits of the Glasford and Banner Formations. Outwash deposits at the base of the Vandalia Member also were included in the lower Glasford/upper Banner aquifer layer because they usually are widely separated from the shallower outwash deposits included in the Glasford aquifer layer. Thus, the elevation of the top of this aquifer layer (figure 12) roughly coincides with the elevation of the top of "marker" bed 2 (Yarmouth Soil-Lierle Clay, figure 9). The elevation of the top of this aquifer layer ranges from above 625 feet in the northeastern part of the study area to lower than 475 feet in the southwestern part of the area. The top is generally inclined toward the west. The lower Glasford/upper Banner aquifer layer appears to be widespread, but it has a calculated thickness generally less than 25 feet (figure 13).

The Glasford aquifer layer includes all of the sand-and-gravel deposits in the upper part of the Glasford Formation, at the base and top of the Radnor Member. The top of this aquifer layer ranges in elevation from more than 675 feet to less than 525 feet and generally descends from east to west across the model area (figure 14). Although this aquifer layer appears to be relatively extensive in the eastern half of the model area, it typically is thin and rarely exceeds 25 feet in thickness (figure 15).

Sand-and-gravel deposits within the Wedron Group were not included in the hydrogeologic model because they occur sporadically and no water-level data are available for them. Nevertheless, their depth and thickness were recorded in the hydrogeologic database so their presence could be used to adjust hydraulic parameters in the ground-water flow model. The digital output from the hydrogeologic model was directly imported into the ground-water flow model.

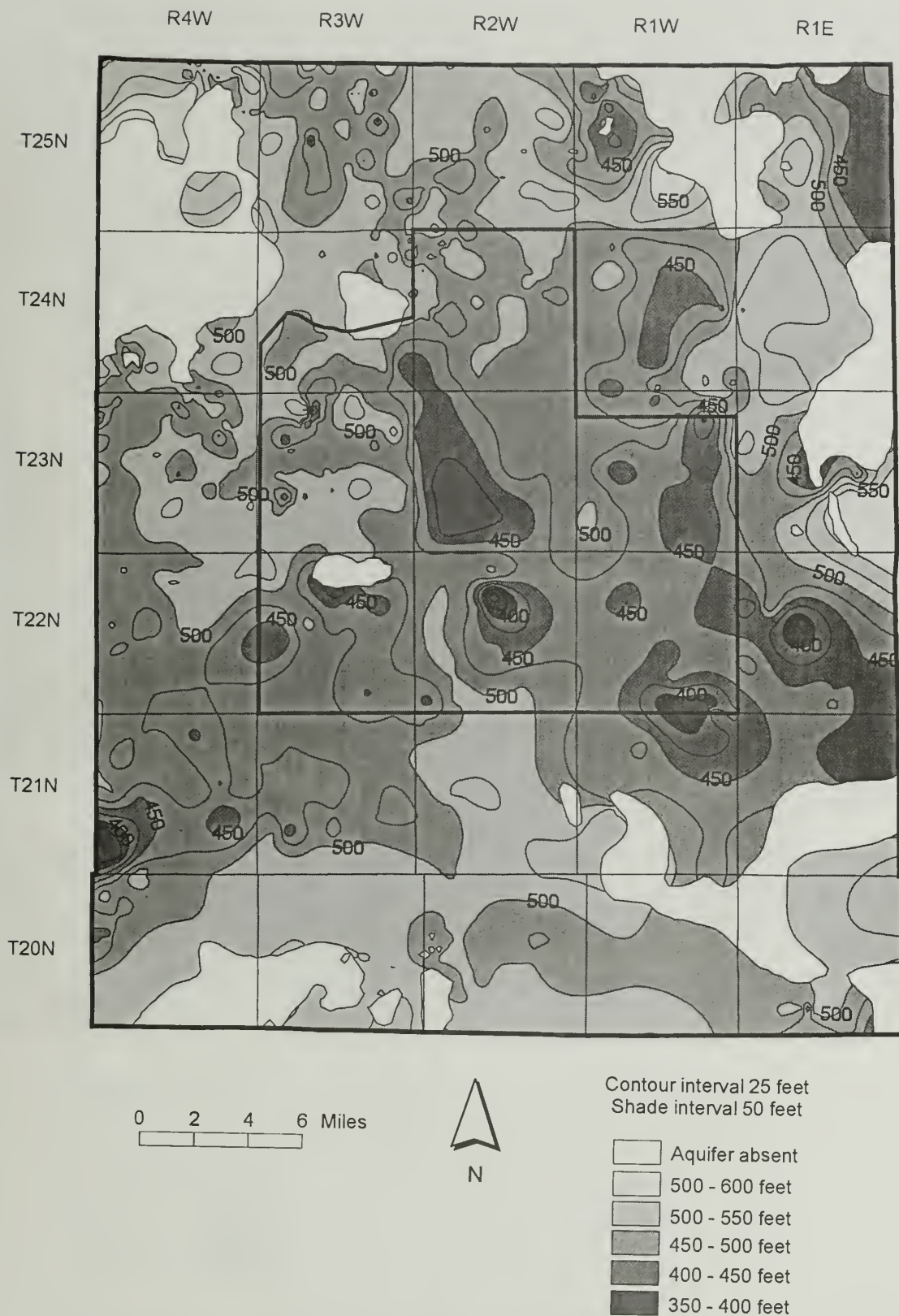


Figure 10. Elevation of the top of the Sankoty-Mahomet Sand aquifer layer

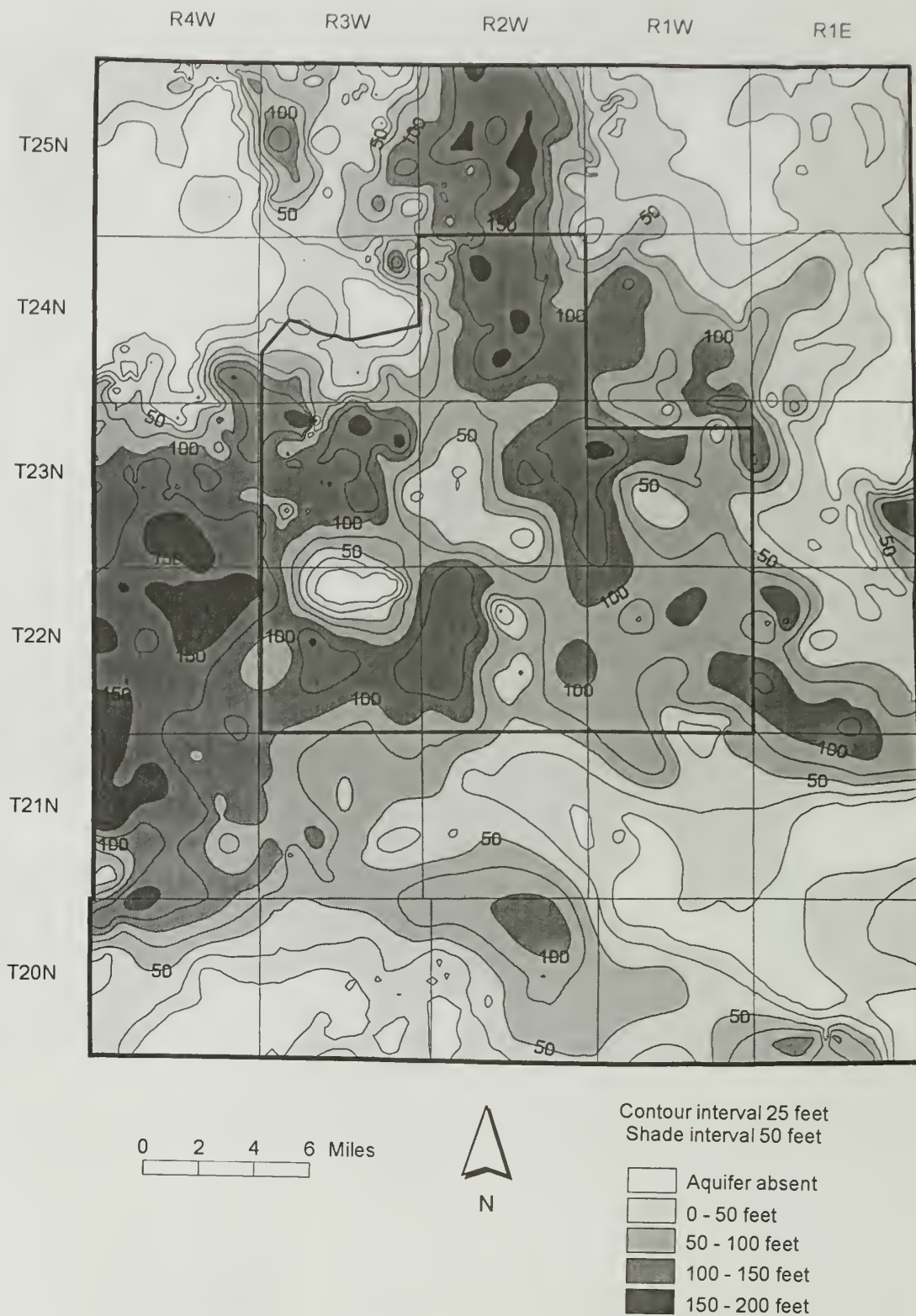


Figure 11. Thickness of the Sankoty-Mahomet Sand aquifer layer

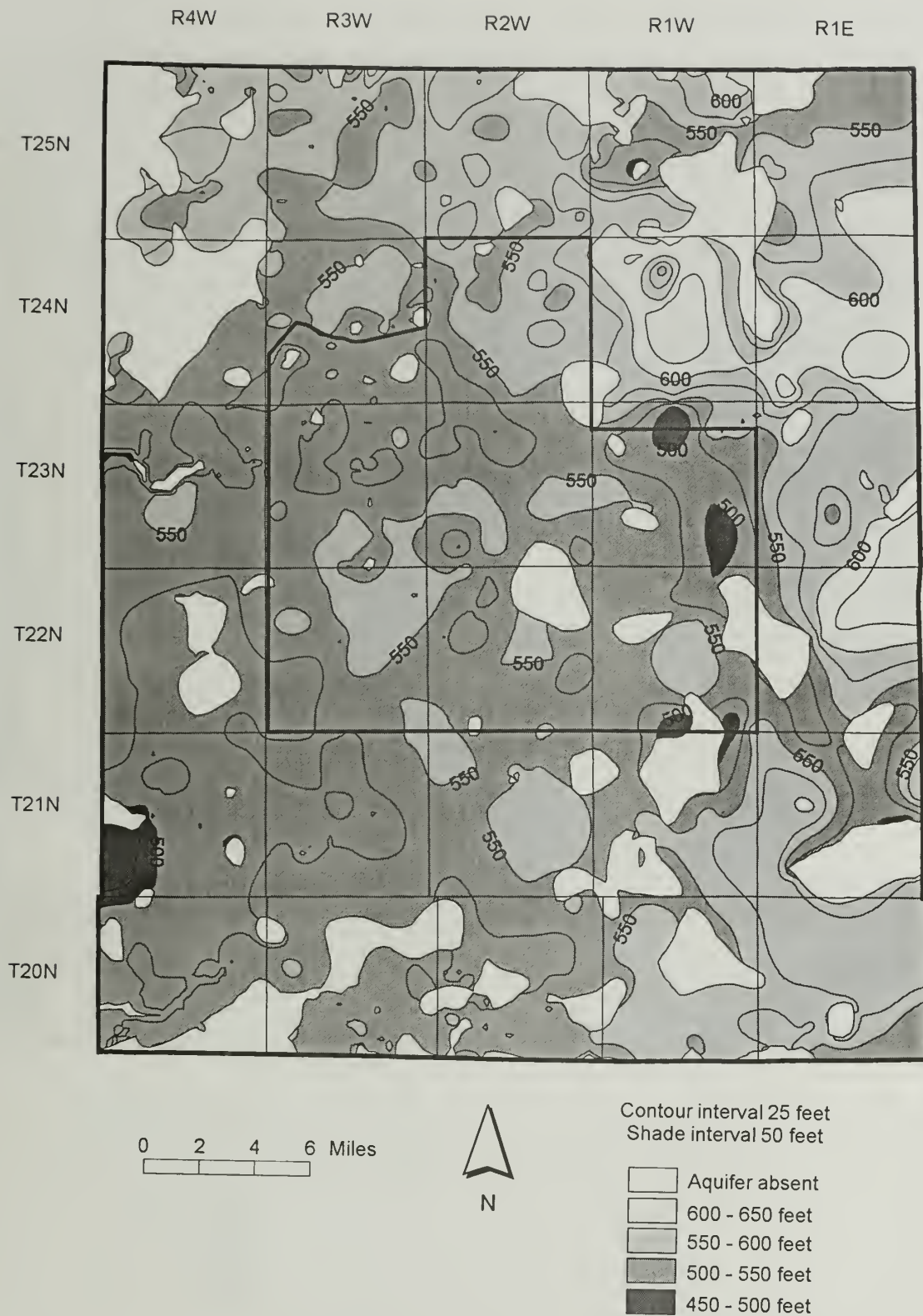


Figure 12. Elevation of the top of the lower Glasford/upper Banner aquifer layer

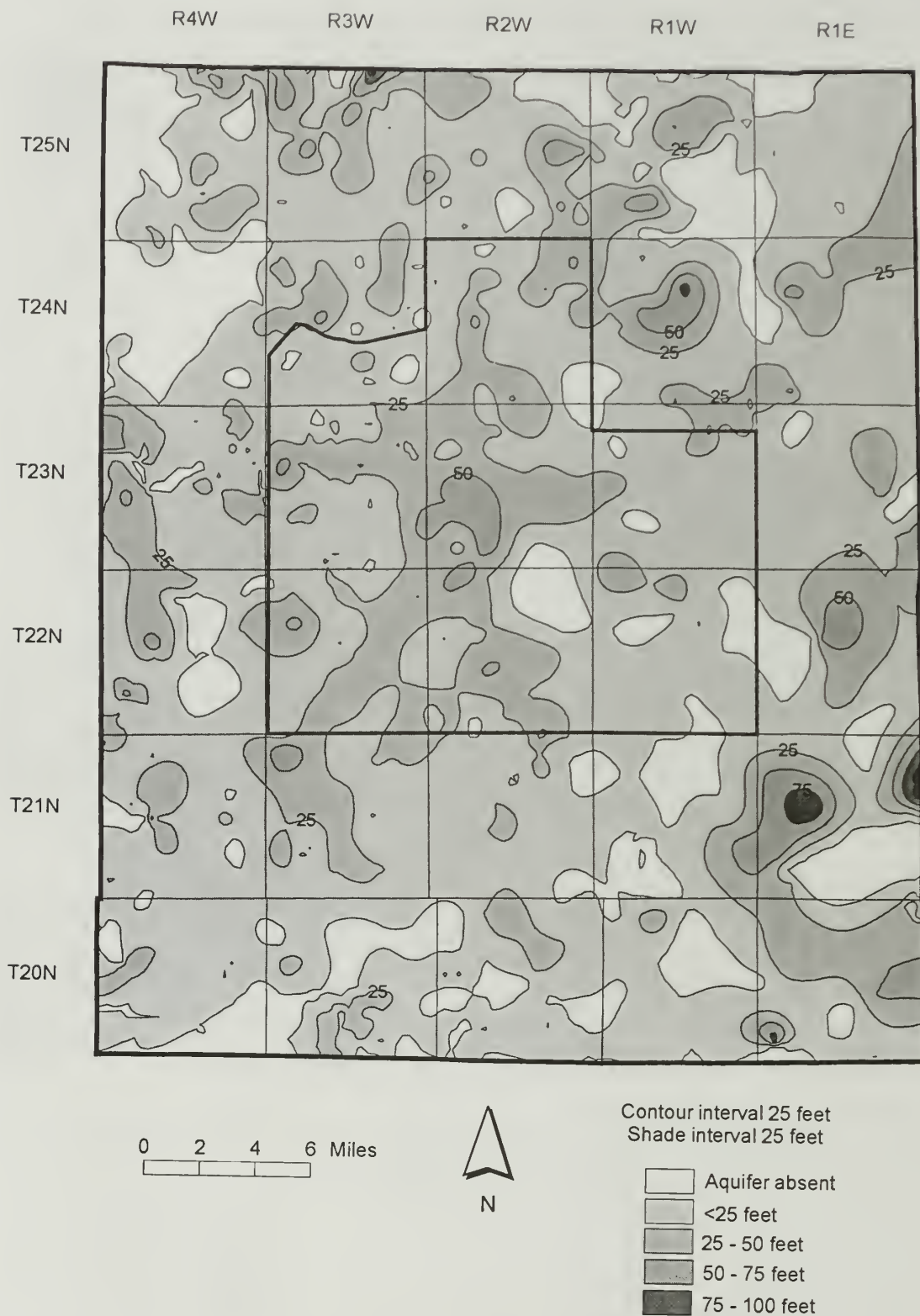


Figure 13. Thickness of the lower Glasford/upper Banner aquifer layer

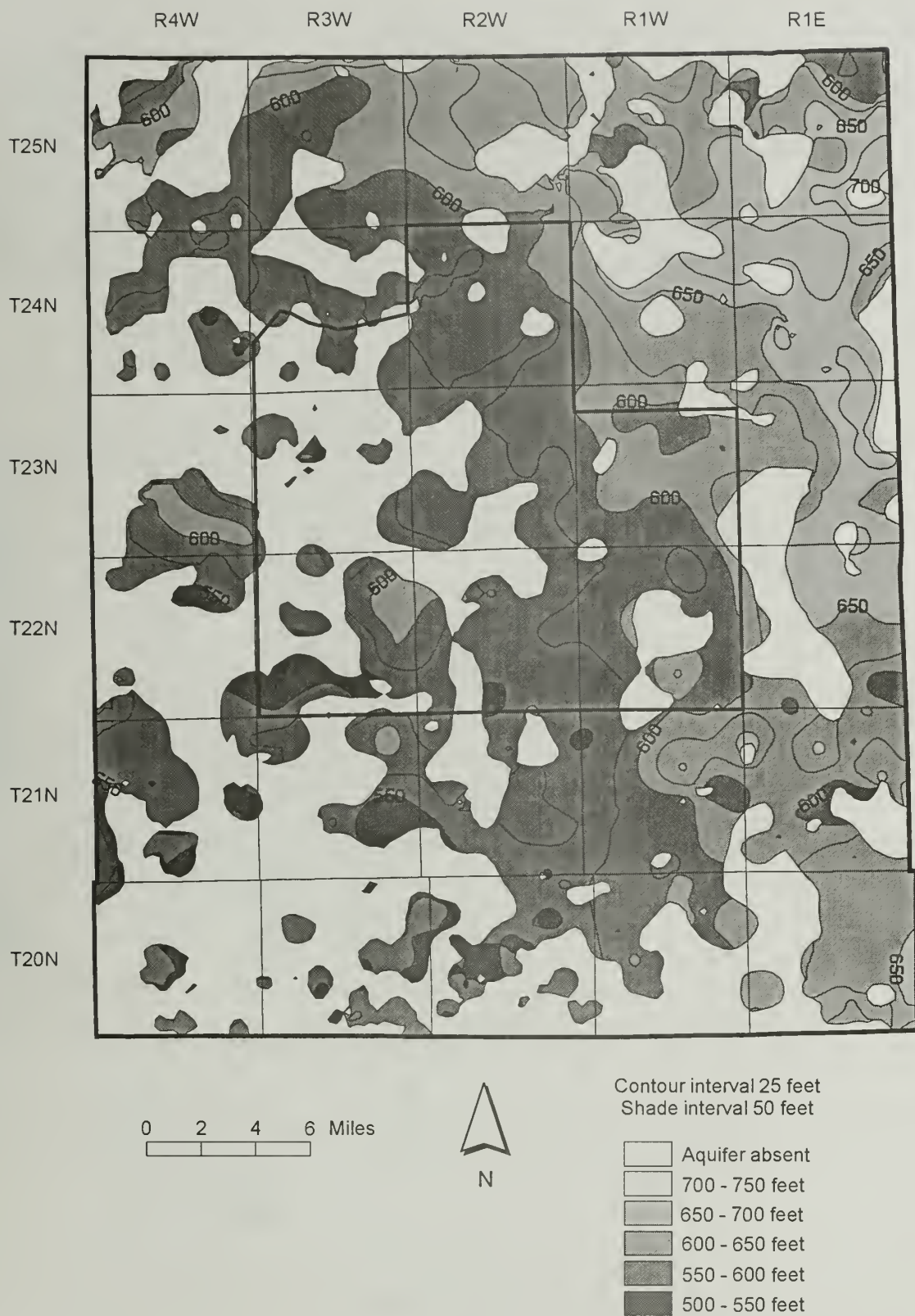


Figure 14. Elevation of the top of the Glasford aquifer layer

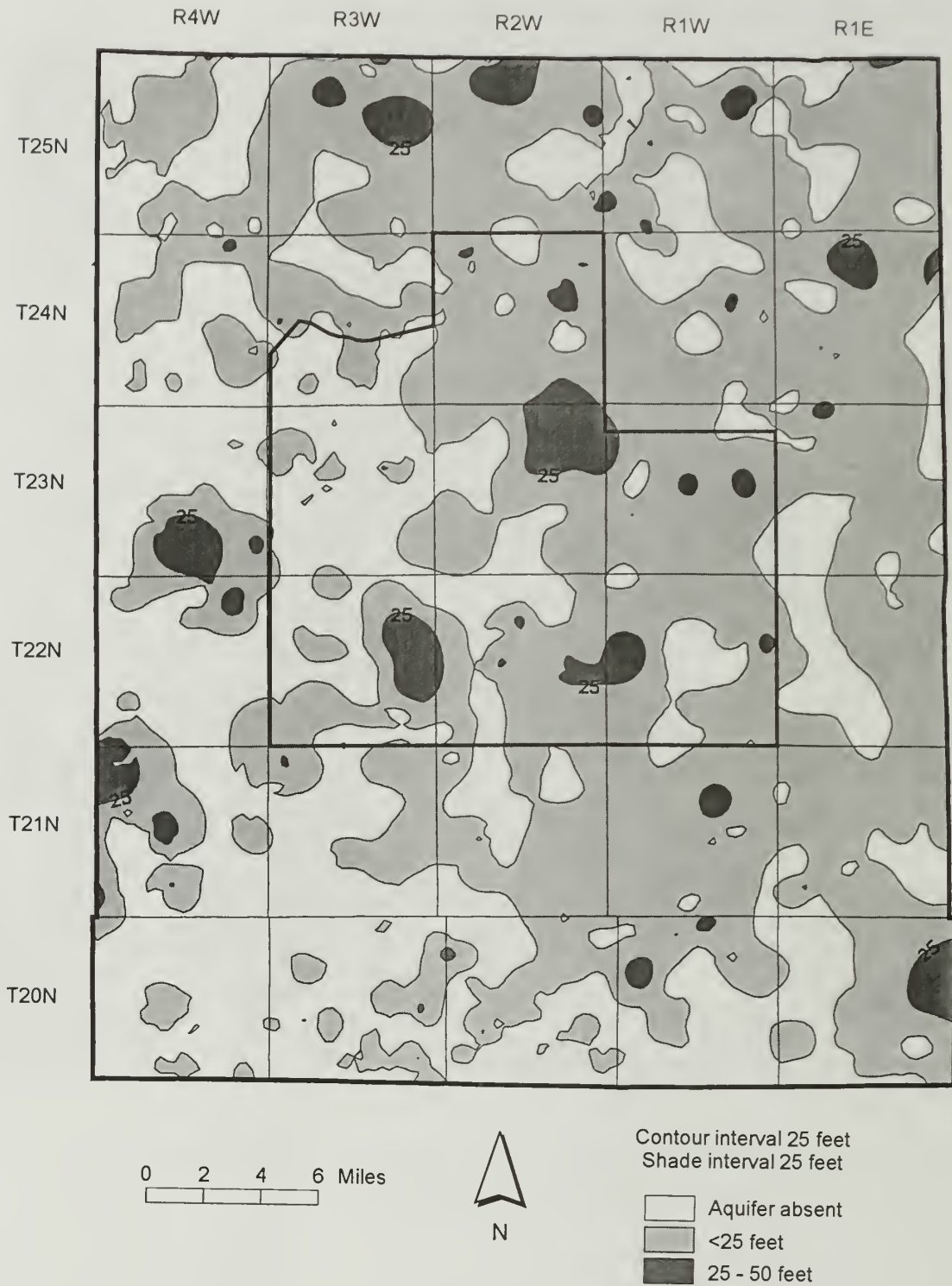


Figure 15. Thickness of the Glasford aquifer layer

HYDROLOGY

A major task of this study was to develop a ground-water flow model that could simulate measured water-levels in wells and actual ground-water flow conditions. These simulations are needed to estimate future water levels in wells and ground-water flow conditions in the study area under various pumping scenarios. The scenarios explored with the ground-water flow model provide insight into the potential for the development and the impacts on the ground-water resources of the Sankoty-Mahomet Sand aquifer.

To develop a representative ground-water flow model, a conceptual understanding of the ground-water flow in the glacial deposits was developed. This understanding was based on the theories of ground-water flow and the conditions in the study area. Hydraulic properties, water-levels, ground-water discharge, and recharge are interdependent. Together, these characteristics determine how ground water flows into, through, and out of an aquifer system.

A confined aquifer occurs between two hydrogeologic units of significantly lower hydraulic conductivity (aquitards). An unconfined aquifer occurs when the upper surface of the aquifer is open to the atmosphere. The potentiometric surface (the level to which water in a well screened in a particular aquifer will rise) in an unconfined aquifer is the water table. Thus, in an unconfined aquifer, the water level is below the top of the aquifer, and the pressure at the water surface is equal to atmospheric pressure. In a confined aquifer, the potentiometric surface and the water level in wells are above the top of the aquifer (artesian conditions) because of hydraulic pressure on the water in the aquifer. If the water level in a confined aquifer drops below the top of the aquifer, the aquifer changes from confined to unconfined conditions. A discussion of how the water-level data were gathered and analyzed is presented later in this report.

Analytical equations that describe ground-water flow to a pumping well are well documented, and an explanation of them was provided in the aquifer characterization report (Herzog et al., 1995b). Aquifer hydraulic properties can be determined by conducting controlled-rate aquifer tests. By monitoring the pumping rate and water-level drawdowns during the test, and by knowing the physical features of the aquifer, these hydraulic properties can be calculated. The aquifer hydraulic properties calculated from the aquifer tests provided the data used to estimate the flow of ground water through the aquifer, determine the rate of vertical recharge from shallower aquifers into the Sankoty-Mahomet Sand aquifer, and estimate the volume of water available in storage in the aquifer. Typically, an aquifer test, such as the two conducted for this project, involves many water-level measurements at multiple observation wells. Aquifer hydraulic properties are calculated separately at each observation well. The aquifer properties data at each well are evaluated together to determine average values for the aquifer that represents the aquifer hydraulic properties for the tested area. If the geologic deposits found at an aquifer test location are similar in lithology to the geologic deposits in other areas in the aquifer, the test results may be assumed to be representative of the properties in the other areas as well. The aquifer characterization report (Herzog et al., 1995b) summarizes all of the previous pumping tests (40+) conducted in the area. These data show that the hydraulic properties determined from two aquifer tests conducted for this study are representative of the hydraulic properties for the Sankoty-Mahomet Sand aquifer in the study area.

An analysis of the water-level data indicates the flow direction of the ground water in the aquifers, identifies areas of recharge and discharge, and provides insights into the seasonal variations of the water level. Water-level data were gathered from several sources for the study. During the study, water levels were measured in approximately 200 wells. Water-level data gathered during several research studies in the areas adjacent to the model area, to the north, east, and west, were used. And, of the 2,278 water-well records provided by drillers, 1,100 included water-level information (depth to water). These data were grouped and plotted to develop maps of the potentiometric surface of each aquifer layer.

River-stage data were correlated with the water-level data to identify areas at which the aquifers are hydraulically connected to the rivers. If hydraulically connected, the two data sets have nearly the same values and/or the changes in level with time will rise and fall together or with a short time lag between the two. River-stage measurements came from two sources. The United States Geological Survey (USGS) maintains two gaging stations on the Mackinaw River near the model area. The one at Congerville is in the northern part of the model area, and the one at Green Valley is west of the model area. These published data are available from the USGS (Coupe et al., 1989). River stage also was measured at three locations along the Mackinaw River in the study area. These measurements, collected in conjunction with the ground-water level measurements, were made using an electric water-level measuring device and a surveyed point on the bridge railing as the reference point.

Regional Ground-Water Flow System in the Sankoty-Mahomet Sand Aquifer Surrounding the Model Area

The regional ground-water flow system in the study area is complex. Data gathered by ISWS and ISGS scientists in the last few years have improved their understanding of the study area, but the recent changes in interpretation also suggest that an understanding of the complete system is continually evolving. The study area and the model area are only part of a much larger regional flow system (figure 16).

Previous research has identified the Sankoty-Mahomet Sand aquifer as potentially capable of supplying large quantities of water for further development in the model area (Kempton and Visocky, 1992; Wilson et al., 1994; Herzog et al., 1995a). Under natural conditions, the Sankoty-Mahomet Sand aquifer is at an equilibrium; the amount of recharge into the aquifer is equal to the amount of discharge from the aquifer. A large withdrawal from the aquifer, such as pumpage from a municipal well, will alter ground-water flow in the aquifer and will lower the potentiometric surface of the aquifer in the vicinity of the pumped well. The degree to which water-levels in wells decline (drawdown) and the areal extent over which decline occurs (cone of depression) depends on several factors. The cone of depression expands until the volume of water flowing toward the well matches the volume withdrawn. Recharge into the aquifer in the area of the cone of depression may be from vertical leakage from above or below, and/or induced horizontal flow toward the pumped well. A cone of depression also may cause natural discharge from the aquifer to be reduced. If these changes equal the pumpage, the potentiometric surface in the aquifer will develop a new equilibrium. If they do not balance, such as when withdrawal from the aquifer is greater than the recharge to the aquifer, water levels in wells will continue to decline, as has been observed in deep sandstone aquifers in the Chicago area (Visocky, 1997). There, the potentiometric surface has declined by nearly 1,000 feet. Recently, as utilization of Lake Michigan water has increased, the potentiometric surface of the deep sandstone aquifer has rebounded by several hundred feet at some locations (Visocky, 1997).

From the model area, the Sankoty-Mahomet Sand aquifer extends eastward into Indiana, northward into the Illinois River Valley, and westward, to the Illinois River. Several segments of the aquifer were studied previously (Anliker and Sanderson, 1995; Burch and Kelly, 1993; Marsh, 1995; Sanderson and Buck, 1995), and water-level data gathered for those studies were combined to develop a regional map of the potentiometric surface (provisional) for the Mahomet and Sankoty-Mahomet Sand aquifers from the Indiana state line to the Illinois River (figure 16).

The water-level data used to construct figure 16 came from six sources:

- Data were collected for this project in McLean, Logan, and eastern Tazewell Counties.
- Data for the area east and north of Champaign-Urbana came from the ongoing ISWS aquifer assessment drilling program.

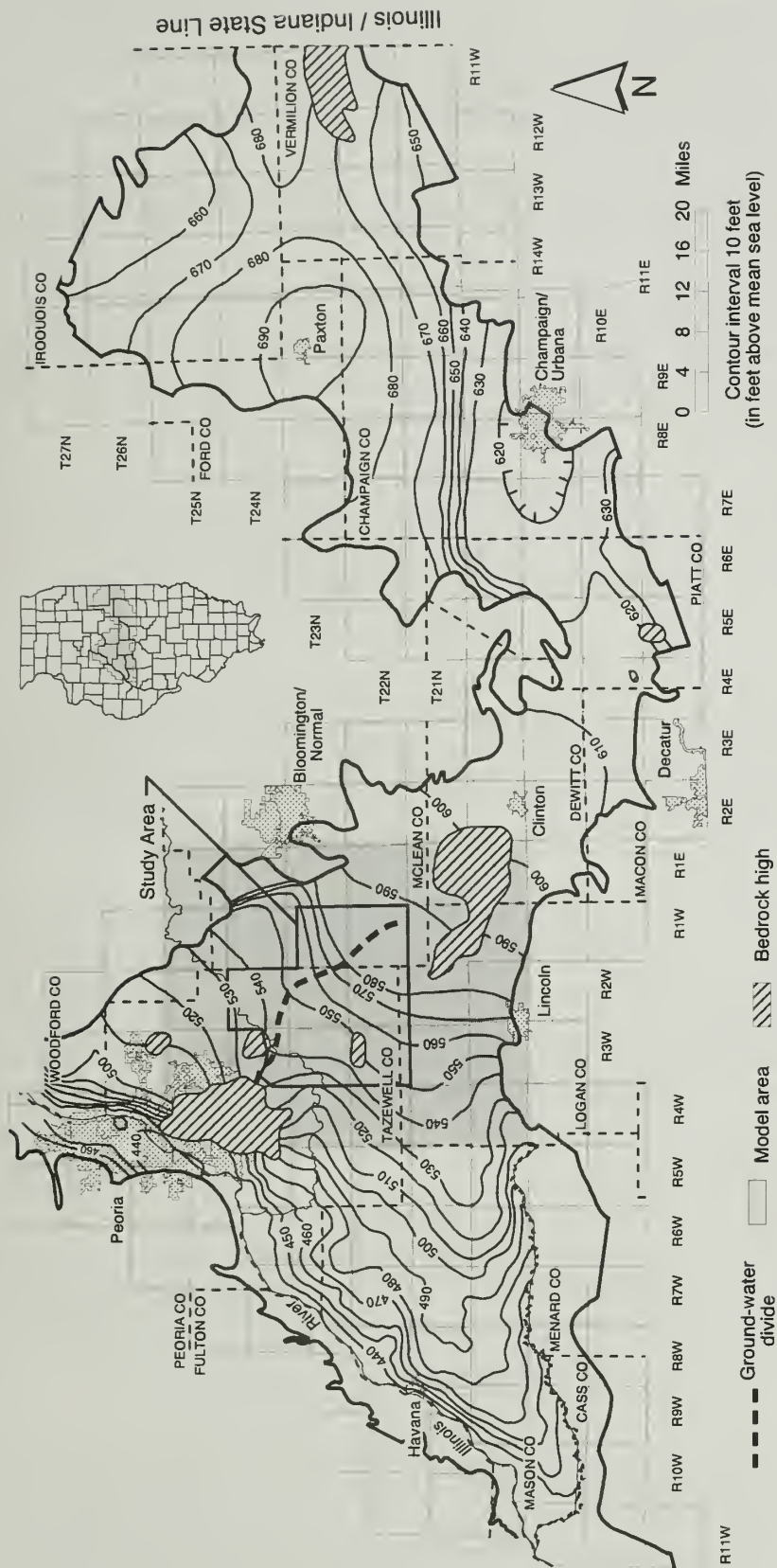


Figure 16. Provisional potentiometric surface of the Mahomet/Sankoty-Mahomet Sand aquifer (numbers indicate contour elevation)

- In DeWitt and Piatt Counties, the ISWS measured water levels in more than 500 private wells during 1994 (Anliker and Sanderson, 1995).
- West of the study area in Mason and western Tazewell Counties, data came from nearly 300 water-level measurements in 1992 and 1993 by Sanderson and Buck (1995).
- Near Morton and East Peoria, water-level data came from a water quality study conducted by Burch and Kelly (1993).
- For the area in northern Vermillion County, for which no water-level data were available from previous research, water-level elevations were estimated from water well records in the ISWS private well database (Marsh, 1995).

The provisional potentiometric surface map in figure 16 shows that ground water generally flows from areas in east-central Illinois westward to the Illinois River. A potentiometric high exists in the Mahomet Sand aquifer near Paxton, in Ford County. The ground-water divide created by this high approximately coincides with the surface water divide between the Ohio and Mississippi River basins.

A cone of depression near Champaign-Urbana (figure 16) is caused by annual withdrawals of roughly 18 mgd from the Northern Illinois Water Corporation's (NIWC) well field. West of the well field, the flow direction is from west to east, toward the cone of depression, the reverse of the regional flow. Ground water in the Mahomet aquifer near the well field is diverted toward the NIWC well field and does not flow west toward the Illinois River.

In the model area, the divergence of flow, shown as a ground-water divide, is an important feature of the ground-water flow system (figure 16). Flow into the model area in the Sankoty-Mahomet Sand aquifer from the east is through the Mahomet Bedrock Valley and the Kenney Valley (figure 17). Flow out of the model area is both northwest toward Peoria and west into Mason and Tazewell Counties. Though the model area has been called the confluence area because of the convergence of the Mahomet and Mackinaw Bedrock Valleys, it is an area of divergence of ground-water flow.

North of the model area, near Morton and East Peoria, the potentiometric surface drops steeply toward the Illinois River. Ground water from the Sankoty-Mahomet Sand aquifer discharges into the Illinois River in this area. Similarly, west of the model area in Mason and western Tazewell Counties, the land-surface elevation is significantly lower, and the top of the Sankoty-Mahomet Sand aquifer is close to or at land surface. The aquifer is unconfined in this area, nearly all recharge comes from direct surface infiltration, and ground water discharges into the Illinois, Sangamon, and Mackinaw Rivers.

Hydrologic Data in the Model Area

Hydrologic data were reviewed and entered into a computer database. Additional field data, collected during the second part of the study, were added to provide insight where uncertainties still existed in the conceptualization of the ground-water flow system. By evaluating where data deficiencies existed and then gathering additional data to fill those deficiencies, a more complete understanding was developed.

Ground-Water-Level Database

The distribution of ground-water-level data was not uniform over the model area or among the aquifer layers. Most data were for the Sankoty-Mahomet Sand aquifer layer. Fewer water-level data were available for the lower Glasford/upper Banner and the Glasford aquifer layers. To supplement these data, river-stage elevations were included in the database for locations at which an aquifer layer was interpreted to be in direct hydraulic connection with a river or stream. These locations were determined to be where the layers were at or

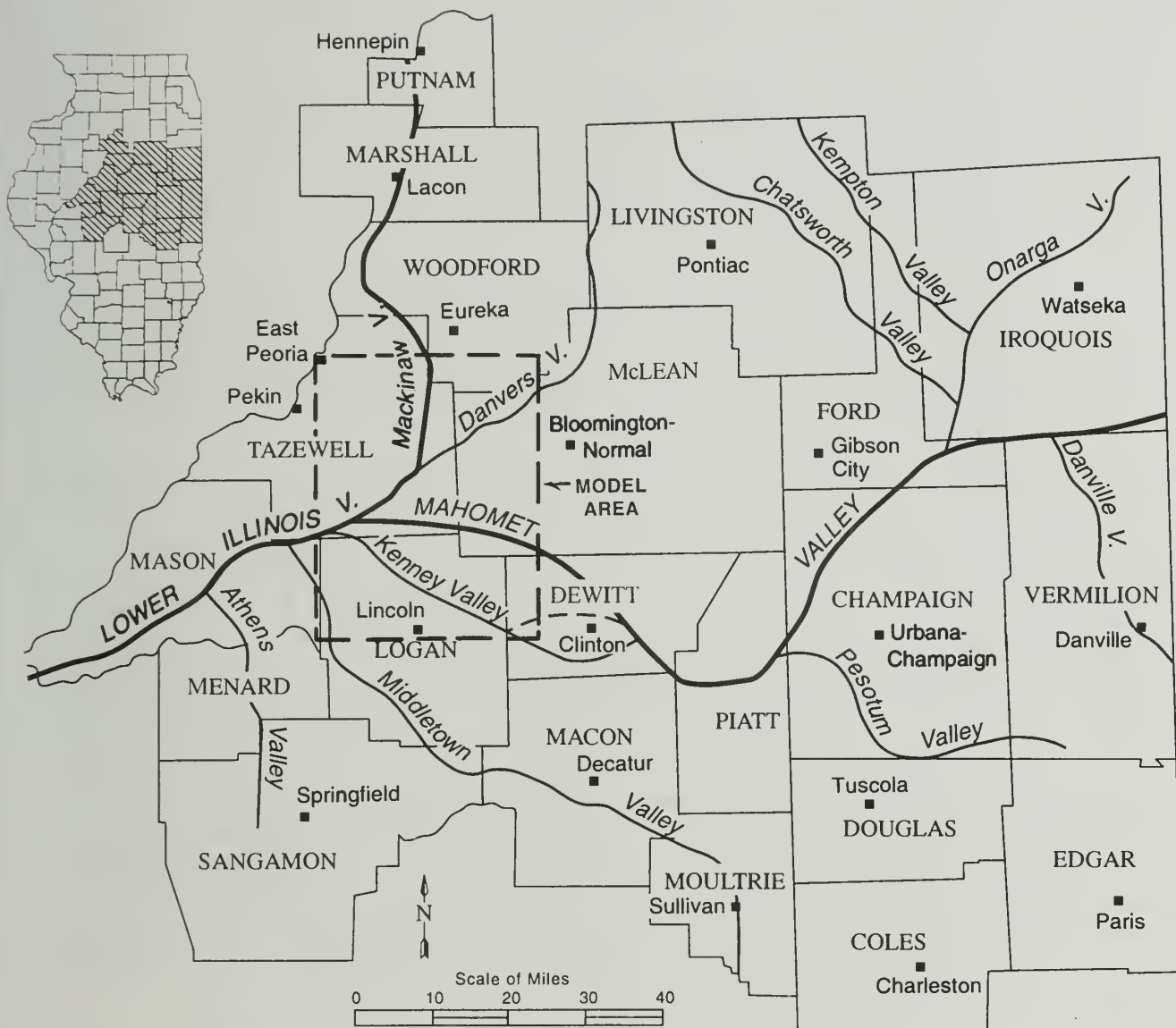


Figure 17. Thalwegs of the major bedrock valleys in central Illinois
(adapted from Water Resources Center, 1997)

near land surface. Because of the direct hydraulic connection, the potentiometric surface of the aquifer layer can be approximated by the river stage at those locations. For the lower Glasford/upper Banner and Glasford aquifer layers, additional water-level data from the aquifer below either aquifer were included in the database at locations with a direct hydraulic connection between aquifer layers. Where the aquifer layers are connected, the potentiometric surface elevation would be virtually the same in both aquifer layers, though there are slight differences due to vertical flow between the two layers. The differences in head will be small compared to the contour interval, and thus were assumed to be equal for creating the potentiometric surface maps.

About 1,500 water-level measurements were available in the model area. Water levels in approximately 400 of the 1,500 available wells were measured during research studies conducted by the ISWS and ISGS since 1992 (Anliker and Sanderson, 1995; Herzog et al., 1995a; Sanderson and Buck, 1995; Wilson et al., 1994), 200 of these 400 wells were measured during the mass water-level measurement made for the aquifer characterization study (Herzog et al., 1995a). The water levels for the remaining 1,100 wells were those reported by the drillers on the water-well records.

The water-level data from drillers' logs were collected over several decades, and differences in water level at a particular well due to seasonal variation or changes from different years were not determined, adding some uncertainty to the data. However, the water-level data collected in the model area since 1992 indicate that the water level did not vary more than 5-7 feet. Therefore, although the water levels from water-well records have some temporal variability, they were assumed to be representative of the water levels for the area. The water-level data collected by the drillers were used to supplement data collected by researchers, and when other data were not available. The water levels from the vicinity of the Normal well field were checked to determine if the measurements were taken before or after the development of the well field.

Stream-Sediment Sampling

To better understand the role of surface-water-ground-water interaction in recharge to and discharge from the aquifers, researchers waded many of the ditches, streams, and rivers in the model area. This effort was undertaken to evaluate the possible locations at which the streams may be directly connected to a sand unit at the land surface. The hydrogeologic layer maps developed for the ground-water flow model and low-flow stream data (Singh and Adams, 1973) were used as a guide to decide which streams to evaluate. Those selected were chosen because either the hydrogeologic data suggested the upper aquifer layer may be close to the land surface or the data suggested that the stream had a gaining reach. Samples of sediment were collected along the stream banks and stream beds using a 2-foot long push tube. Samples were taken as much as 5 feet into the stream bank and stream bed when the sampling hole remained open. The type of material collected in the sampler was used to determine if a particular stream was potentially connected hydraulically to an upper sand unit. The samples fell into three classes: (1) clay stream banks and a clay stream bed, (2) clay stream banks and a sand stream bed, and (3) sand stream banks and a sand stream bed.

Sand in the stream bed could have been deposited by the stream (bedload) or could be from a sand unit in the Wedron Group, or any underlying unit. Although this method is not straightforward because thickness and continuity are not known, this information was useful. In conjunction with the maps of the aquifer layers, the information allowed researchers to determine areas at which an aquifer layer in the ground-water flow model might be hydraulically connected to the stream. This was especially helpful at locations where the USGS 7.5-minute topographic maps did not adequately represent the stream elevation, or where geologic data were sparse. At some bridges, the stream bed elevation, based on the mapped elevation of the bridge, was as much as 15 feet below the mapped land-surface elevation. Because land-surface elevations from 7.5-minute topographic maps were used to create the top of the upper layer in the ground-water flow model, elevations at these locations were corrected in the model. When the data from geologic logs indicated that the sand at or near land surface was not

from an aquifer layer of the ground-water flow model but a Henry Formation sand, these data were used to identify reaches of streams in which infiltration may be greater than those reaches with clay bottoms. This information was useful for setting up river node parameters in the ground-water flow model, such as vertical hydraulic conductivity, when no other data were available.

Potentiometric Surfaces and Ground-Water Flow

Potentiometric surface maps were prepared for each of the three aquifer layers in the ground-water flow model (figures 18-20). Flow in an aquifer is perpendicular to the contour lines of equal hydraulic head. Table 2 presents the number of data measurements used in the construction of the potentiometric surface maps. The potentiometric surface maps represent the water-level elevations measured in wells in the aquifer layers. The data were of three types: (1) water-level elevations from mass measurements, (2) water-level elevations taken from historic well records, and (3) river-stage elevations for areas in which the aquifer and river were hydraulically connected. The potentiometric surface maps created for the ground-water flow model are based on the data collected during the fall 1994 mass measurement conducted for this project. For the limited time frame that adequate water-level data from wells were available (1994 and 1995), the data for 1994 best represented the "normal" or "average" value of the water levels. A review of the local climate data for the area from 1993 to 1997 indicated that 1994 and 1995 were both only about 3 inches from the "average" annual precipitation; the other years varied more. In 1995, though a near normal precipitation year, the distribution of rainfall during the year was not normal; May rainfall was more than 8 inches above the May monthly "average."

Each of the three aquifer layers is absent in some part of the model area. The two upper aquifer layers were created by combining separate geologic units into two composite hydrogeologic layers. Simplifying the hydrogeology in this way for the ground-water flow model created aquifer layers that are more continuous than actually exist in the aquifers of the study area. The potentiometric surface of these upper aquifer layers were mapped as being equally continuous, when in reality they are not. Because the upper two aquifer layers are thin compared to the Sankoty-Mahomet Sand aquifer layer, the effect of this simplification in the ground-water flow model should be minimal. In some areas the confining layer between two aquifer layers is absent. At these locations, the aquifer layers are physically connected, making the mapped surface between them somewhat arbitrary, and they act as one aquifer unit.

To construct the potentiometric surface maps in a way that took advantage of what was known about the connections between the aquifers, the potentiometric surface map of the Sankoty-Mahomet Sand aquifer layer was created first. This aquifer has the highest transmissivity and exerts the greatest influence on water levels in the overall ground-water flow system. In addition, substantially more water-level data were available for the Sankoty-Mahomet Sand aquifer than for the shallower aquifers. Therefore, in areas that the Sankoty-Mahomet Sand aquifer and the lower Glasford/upper Banner aquifer were hydraulically connected, data from the completed Sankoty-Mahomet Sand aquifer layer potentiometric surface were used to supplement the data

Table 2. Number of Data Points Used to Produce Each Potentiometric Surface Map

<i>Aquifer layer</i>	<i>Number of mass water-level measurements</i>	<i>Number of water-level values from well records</i>	<i>Number of stream level measurements</i>
Glasford	15	37	52
Lower Glasford/upper Banner	37	34	11
Sankoty-Mahomet Sand	179	62	4

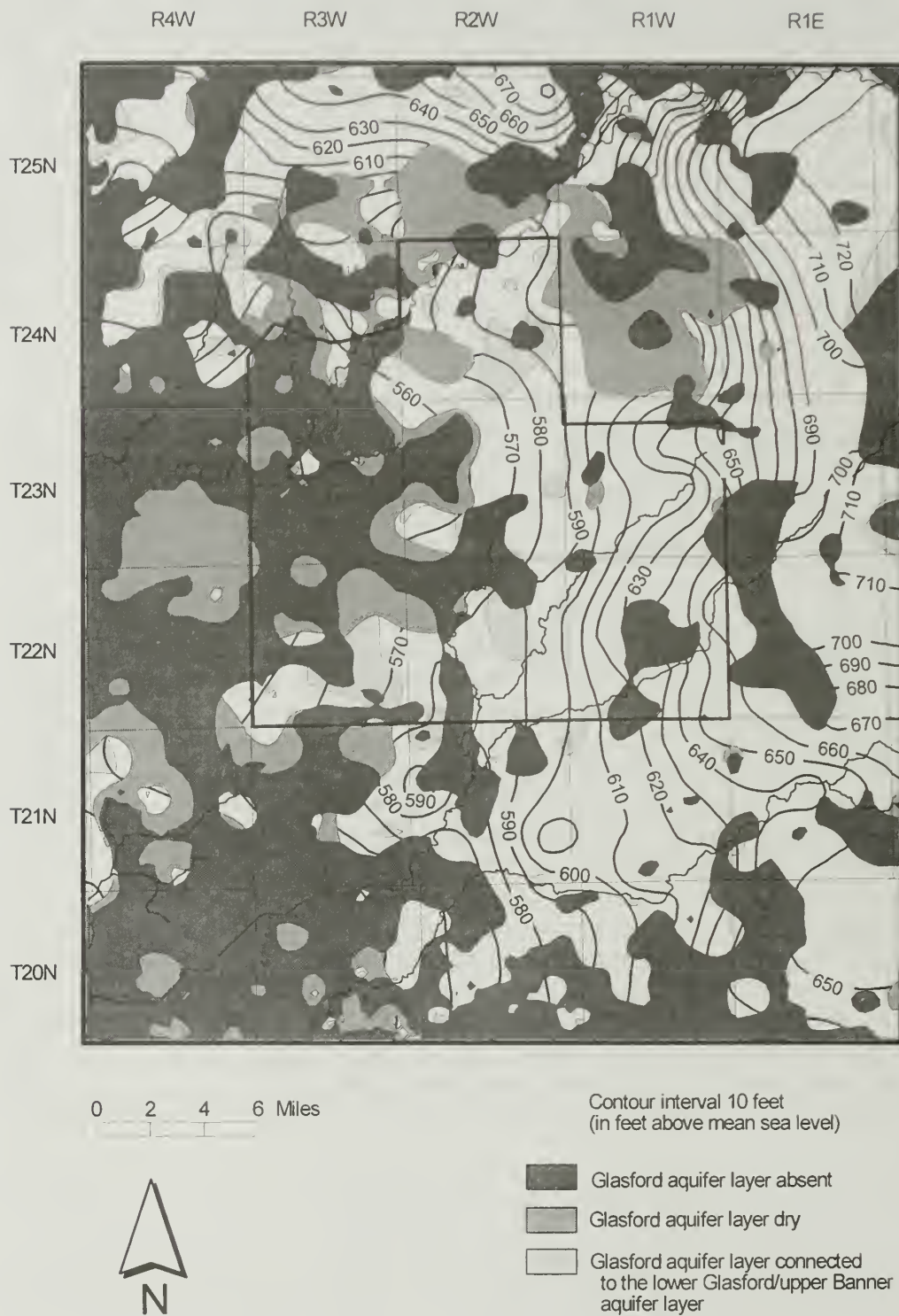


Figure 18. Potentiometric surface of the Glasford aquifer layer in the model area (numbers indicate contour elevation)

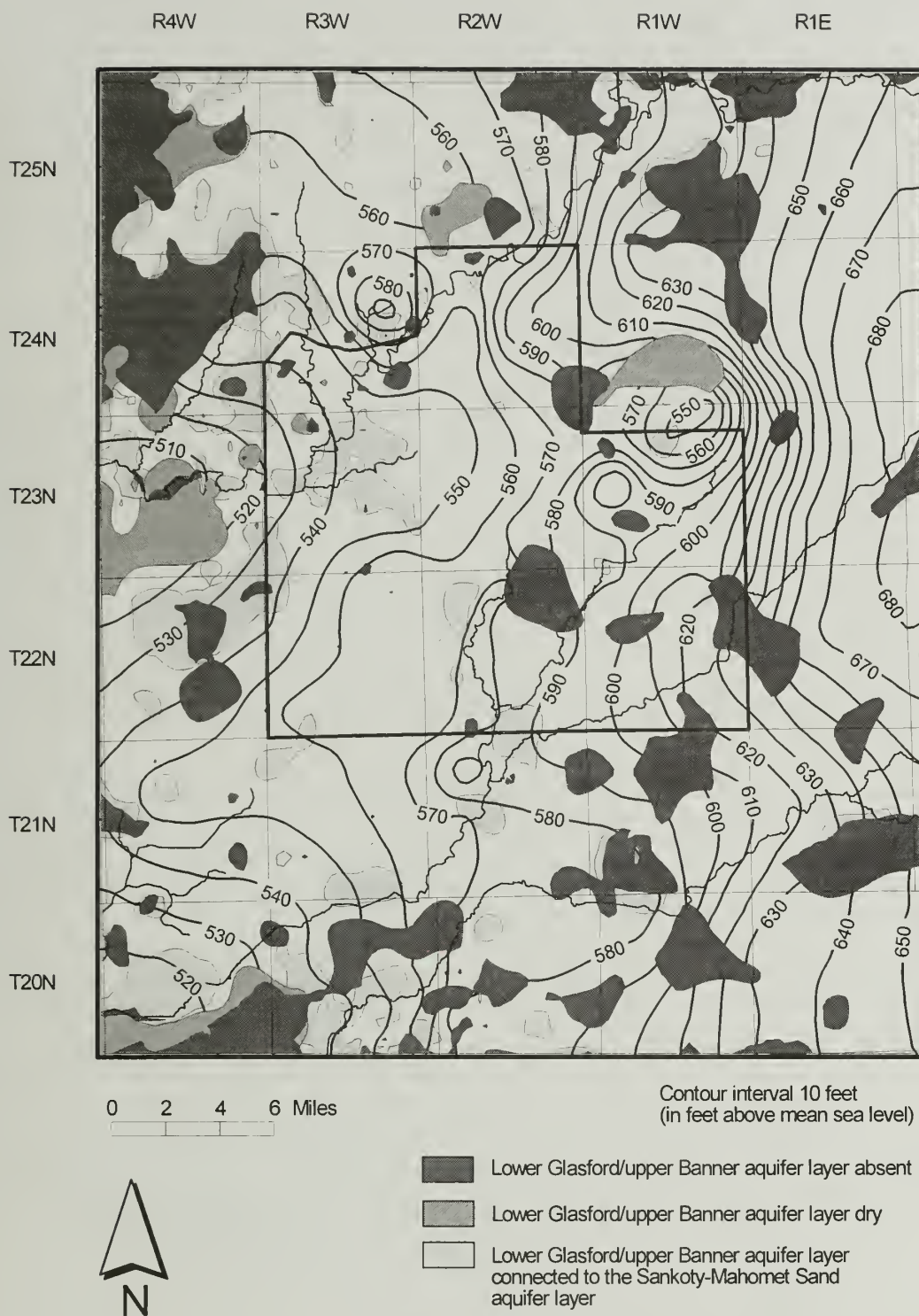


Figure 19. Potentiometric surface of the lower Glasford/upper Banner aquifer in the model area (numbers indicate contour elevation)

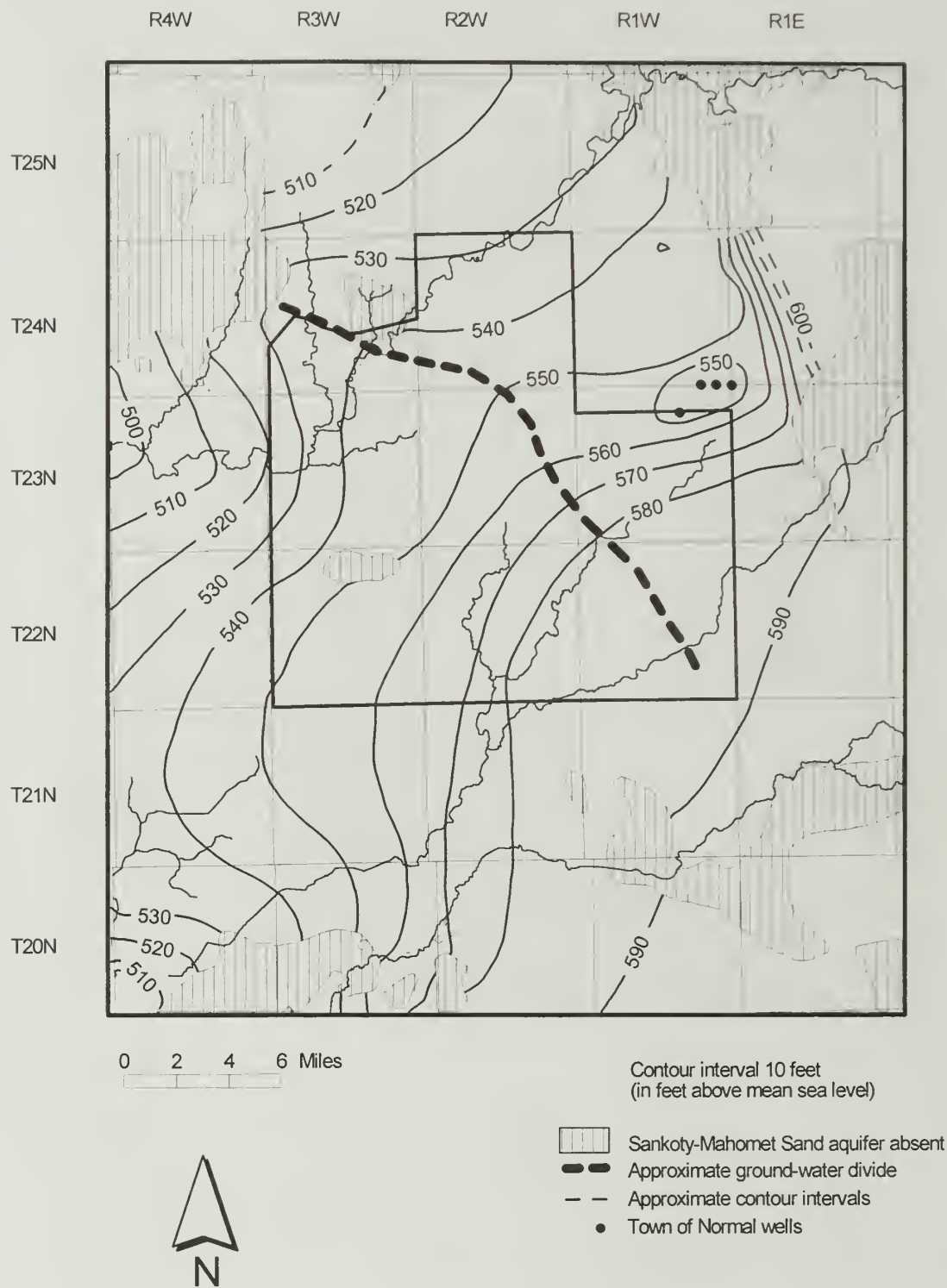


Figure 20. Potentiometric surface of the Mahomet/Sankoty-Mahomet Sand aquifer in the model area (numbers indicate contour elevation)

available for the lower Glasford/upper Banner aquifer layer. Similarly, the Glasford aquifer potentiometric surface map was constructed using some potentiometric surface elevations from the lower Glasford/upper Banner potentiometric surface where a connection existed between those layers.

The development of all three potentiometric surface maps was an iterative process. Development of the potentiometric surface maps of the upper two aquifer layers involved several iterations due to the presence of unsaturated areas. These unsaturated intervals commonly occur along the stream valleys and may occur where the aquifer layer is higher than the surrounding areas of the aquifer layer. These areas were identified by subtracting the potentiometric surface maps from the aquifer bottom maps. The water levels then were contoured again to reflect these new no-flow boundaries.

Glasford Aquifer Layer

The range of elevations on the potentiometric surface map for the Glasford aquifer layer (figure 18) help support the geologic finding that the Glasford aquifer is discontinuous. Reported water levels varied as much as 150 feet over short distances from water-well records, suggesting that individual measurements were from wells in isolated units, not in an extensive, continuous aquifer. The discontinuous nature of this aquifer limits its ability to supply ground water in large quantities. Although several hundred private wells use the Glasford aquifer, these withdrawals are limited. Wells completed in some parts of the Glasford aquifer can be significantly impacted by pumpage from the lower aquifers because of hydraulic connections between them. The Glasford aquifer acts as a source bed for leakage to the lower aquifers in these instances. South of Danvers, in the area of the Normal well field, the Glasford aquifer is unsaturated. The Glasford aquifer is hydraulically connected to regional streams at several locations at which the streams have downcut through the uppermost aquitard layer. The areas of connection are distinguishable on figure 18 by the contours that point upriver, indicating ground-water flow is toward the river.

The patterns of the contour lines identify certain features of the flow system. In areas with the contour lines close together, the hydraulic gradient is steeper. This occurs in the northeast portion of figure 18; it can be caused by a decrease in the hydraulic conductivity of the aquifer layer in a specific location or a reduction in thickness of the aquifer. Water levels will "build up" behind this area, increasing the difference in water-level elevation across it. The closed contours represent areas where the aquifer is either gaining or losing water. In many cases, the closed contours are local features and are not readily explainable. In these instances, they are due to a local phenomena for which supporting data are not available.

The potentiometric surface map of the Glasford aquifer layer generally parallels the trend of land-surface elevation. The highest water levels are near an elevation of approximately 720 feet, and the lowest are near an elevation close to 550 feet. The aquifer layer is generally unsaturated along the Mackinaw River. The unsaturated areas are at locations at which the bottom elevation of the aquifer layer rises above the level of the nearby potentiometric surface in the aquifer. There also are areas in which the aquifer is partially saturated, or unconfined. Confined areas also exist in the Glasford aquifer when the potentiometric surface is above the top of the aquifer.

Lower Glasford/Upper Banner Aquifer Layer

The lower Glasford/upper Banner aquifer layer is more continuous across the model area than the Glasford aquifer layer. This promotes more regionally continuous ground-water flow within the lower Glasford/upper Banner aquifer layer. Water levels measured in wells in this aquifer layer tend to be representative of the regional aquifer layer and not of isolated sand units as usually is the case with the Glasford aquifer layer. The highest water levels in the lower Glasford/upper Banner are at an elevation of about 690 feet, and the

lowest elevations, which occur where the Mackinaw River leaves the model area, are at an elevation of less than 510 feet. The water-level elevations average about 20 feet less in elevation than the Glasford aquifer layer.

The depression shown by the closed contour lines T23N, R1W on figure 19 indicates a discharge from the aquifer layer. This depression occurs because of a localized area of hydraulic connection between the Sankoty-Mahomet Sand aquifer and the lower Glasford/upper Banner aquifer layer. Pumpage from the Normal well field has lowered the potentiometric surface of both aquifers in this area because lowering the water level in the Sankoty-Mahomet Sand aquifer increases the gradient between the two aquifers, which increases the leakage of ground water into the Sankoty-Mahomet Sand aquifer from above. This hydraulic connection and its effects on private wells from pumpage at the Normal well field were well documented in previous studies (Richards and Visocky, 1982; Kempton and Poole, 1985).

Sankoty-Mahomet Sand Aquifer Layer

The Sankoty-Mahomet Sand aquifer is the largest aquifer resource in the model area. It is the only aquifer in this area that can potentially support the development of a large ground-water supply. The potentiometric surface map for the Sankoty-Mahomet Sand aquifer layer (figure 20) indicates that ground-water flow into the model area is from the Mahomet Bedrock Valley east of the model area. The contribution into the model area by the aquifer in the Danvers Bedrock Valley appears to be minimal. Ground water in the aquifer flows both west and north out of the model area. A ground-water divide trends from northwest to southeast (figure 20).

Flow lines derived from the potentiometric surface map (figure 20) of the Sankoty-Mahomet Sand aquifer suggest that the Mackinaw River and Sugar Creek are zones of discharge for the aquifer in the western portions of the model area. The elevation of the top of the aquifer layer, as mapped in figure 10, indicates that the top of the layer is at land surface along reaches of these streams on the west side of the model area; this supports the hypothesis that these streams are hydraulically connected to the Sankoty-Mahomet Sand aquifer and provide an outlet for the ground water flowing from this aquifer.

The effects of ground-water pumpage in the Sankoty-Mahomet Sand aquifer at the Normal west well field are evident on figure 20. The pumpage at the well field lowers water levels enough that the potentiometric surface steepens and lowers toward the well field. Flow in the aquifer in the surrounding area is toward the well field. Because the lower Glasford/upper Banner aquifer layer is hydraulically connected to the Sankoty-Mahomet Sand aquifer near the Normal well field, pumpage of the lower aquifer induces flow from the lower Glasford/upper Banner aquifer layer and lowers water levels in that aquifer (figure 19).

Ground-Water-Level Fluctuations

During the aquifer characterization phase of this project, 28 observation wells were installed in 1993 to provide water-level data at selected locations throughout the study area. In addition, the ISWS had installed 15 observation wells in the model area in 1992 as part of an earlier study (Wilson et al., 1994). The locations of the observation wells are shown in figure 21. Regular, periodic measurement of these wells provided a record of water-level fluctuation for each well. This is important because relationships between the ground-water fluctuations and other hydrologic influences (e.g., pumpage, river stage, and rainfall) can be interpreted from these data. It would be more revealing of the characteristics of the ground-water flow system to have had water-level data over a longer period of record that included more extreme hydrologic events, such as the drought of 1988.

Noting the importance of developing a long-term record of changes in water levels in the area, the LRWPSC provided the resources to maintain this valuable network of observation wells beyond the end of the

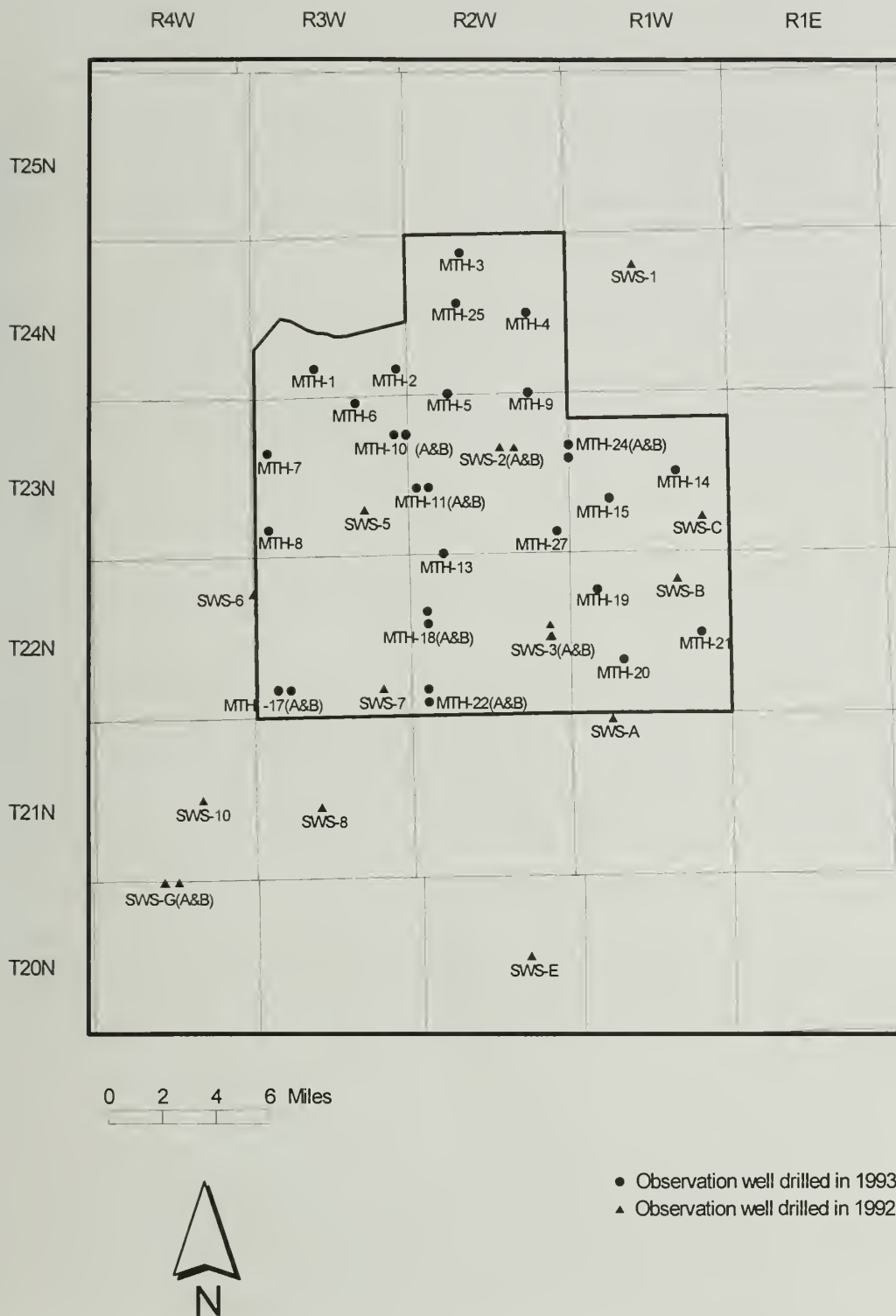


Figure 21. Location of the observation wells

project. The observation well installed in test hole MTH-27 during this second phase of the project was added to the observation well network, and all 44 wells are now measured quarterly. A report to the LRWPSC, which summarized the water-level measurements, included the water-level hydrographs (appendix D) for the observation wells through the November 8, 1997, measurement.

Interpretation of the water-level hydrographs provided many insights into the flow direction, hydraulic gradient, and influence of rivers and pumpage on flow and hydraulic head in the aquifers. For example, the hydrographs for two observation wells finished in the Sankoty-Mahomet Sand aquifer, MTH-1 and MTH-6 (appendix D), both within a mile of the Mackinaw River, appeared to have water-level fluctuations that were similar to the water-stage changes in the river (RVR-6) for the May 1995 water-level high (appendix D). Although most of the observation well hydrographs in the model area had this peak in water level, MTH-1 and MTH-6, in particular, more closely matched the time of the peak in the RVR-6 hydrograph, suggesting a more direct connection. Some wells in the Sankoty-Mahomet Sand aquifer, such as SWS-A (appendix D), had a more delayed and damped peak, indicating that the influence of the river is reduced as the distance from the river increases. Other water-level fluctuations, such as the water-level declines observed in wells near the 30-day aquifer test conducted during April and May of 1995 at Mackinaw, also were evident on the hydrographs. MTH-3, MTH-4, and MTH-25 (appendix D) all showed responses to pumping during the test that are unique to those hydrographs. Other observation wells, further from the test site, showed no observable effects of the pumping test. Hydrographs from other wells screened in shallower aquifers, such as the one for SWS-2S (appendix D), suggest that the aquifer screened by this well may be hydraulically isolated from the others in the area. In this specific case, the well was screened in the lower Glasford/upper Banner aquifer layer for which little observation well water-level data were available for comparison over the model area.

The relationship of the water-level fluctuations to changes in river stage has been interpreted for this study to indicate two important aspects of the ground-water flow system. The Sankoty-Mahomet Sand aquifer is hydraulically connected to the water table and to the unconfined areas west of the study area. The hydrogeologic data support this interpretation, as do the potentiometric surface maps (figures 16 and 20), which indicate that flow in the Sankoty-Mahomet Sand aquifer is to the west and toward the Mackinaw River and Sugar Creek. The May 1995 increase in water levels in wells in the Sankoty-Mahomet Sand aquifer was rapid, a pressure response of the confined aquifer to changes in other parts of the ground-water flow system. This response occurred because (1) of an increase in weight of the materials above the aquifer as they became more saturated, which increased the hydrostatic pressure; (2) the natural outflows of the Sankoty-Mahomet Sand aquifer in the western portion of the model area also experienced a rise in water levels. The increased water levels at the outflow of the Sankoty-Mahomet Sand aquifer propagated back into the flow system. This observed response is an elementary physics relation that historically has been applied to ground water in discussions of aquifer response to tidal influences (Todd, 1980; Hubbert, 1940).

Ground-Water Recharge and Discharge

Ground-water recharge is a general term for the addition of water to an aquifer by external sources, such as infiltration of precipitation, leakage from overlying or underlying aquifers, and seepage from streams. Conversely, ground-water discharge is a general term for the water that leaves an aquifer through mechanisms such as seepage into streams, withdrawal by production wells, and leakage to overlying and underlying aquifers. If only a portion of an aquifer is being studied, recharge and discharge also can include lateral inflow and outflow across the boundaries of the study area.

Quantifying each of the recharge and discharge terms for a flow system requires the formulation of a hydrologic budget in which recharge is balanced by discharge plus changes in storage. Many of the components

in a hydrologic budget, such as pumpage, streamflow, and precipitation, can be measured directly; others, such as ground-water inflow and outflow from other aquifers and changes in storage, must be calculated from hydraulic properties and water-level information. Still others, such as the infiltration from precipitation, often are estimated by indirect means. The number and precision of components that need to be quantified is dependent on the scale of the problem. For example, a detailed study of a farm plot would require collection of flow values for all the processes occurring near land surface, such as plant uptake and discharge to drainage tiles, to estimate infiltration on the plot. Meanwhile a study of a regional aquifer is concerned about the infiltration downward from the land surface and into the aquifer, and infiltration probably would be estimated as an average value for the entire study area, thereby lumping the individual components together.

The recharge and discharge components are interdependent, so quantifying individual components is difficult. For example, ground-water withdrawal by production wells is more accurately known than the other components of discharge, if the wells have an accurate flow meter. In contrast, streamflow measurements can be readily obtained; but, because regional aquifers tend to discharge into the larger streams, the fractional increase in flow attributed to ground water may not be large enough to quantify accurately. The accuracy of estimates of the ground-water inflow and outflow from external areas and of the amount of leakage between aquifers depends on a reasonable understanding of the water-level data, of the geologic relationships, and of the hydraulic properties.

The infiltration of precipitation into an aquifer is the most difficult component to measure directly in the field. A number of methodologies have been used to try to obtain this value. Most methods require a detailed data-gathering effort, either over a large area with a large number of observation wells over a long period of time, or at a relatively small site at which a computed value may not be extrapolated to the rest of the system. In either case, significant effort and resources must be expended. More commonly, infiltration is estimated by indirect means such as flow-net analyses, ground-water runoff determinations, or determination of the mixing ratios of chemical tracers. The classical method of hand drawing flow nets has been replaced by ground-water flow models that generate more detailed and sophisticated flow nets which account for spatial variations in aquifer hydraulic properties and geometries and for the influences of other recharge and discharge components.

Previous Estimates

A commonly used method of formulating a ground-water budget assumes that the baseflow of large streams comes from ground-water discharge, also known as ground-water runoff. The baseflow is calculated using various hydrograph separation techniques or a flow-frequency plot. Baseflow is then divided by the upstream area of the drainage basin to derive an average recharge value over the basin. A recharge value calculated by this method has several important limitations that apply to this study. First, the amount of baseflow in a stream is greatly affected by the regulation of flow by upstream dams and by the discharge of treated wastewater. Second, there is a significant amount of water flowing out of bank storage or through shallow sands and drain tiles that never participates in the regional ground-water flow system. Third, the boundaries of ground-water basins generally do not coincide with the surface-water basins. Together, these limitations tend to produce a recharge budget that is erroneously high. Referring to this value as the recharge to a regional aquifer from the net infiltration produces an additional error, because the other recharge and discharge components, such as upstream river leakage or withdrawal by production wells, are not taken into account.

Walton (1965) estimated recharge budgets from the baseflow of two streams in the study area for years of normal precipitation. Data from the gaging station operated by the USGS on Sugar Creek near Hartsburg produced a ground-water runoff value of 0.001 feet per day or ft/d (207,000 gallons per day per square mile or gpd/sq mi, 4.35 inches/year); the data from the Mackinaw River near Green Valley had a value of 0.00081 ft/d.

Using data from other streams in the region, Walton (1965) generalized ground-water runoff in the northern part of the study area to range between 0.00062 and 0.00093 ft/d, and in the southern part of the study area to range between 0.00093 and 0.0012 ft/d. O'Hearn and Gibb (1980) revisited the Sugar Creek station data, but used a statistical approach in selecting the flow data to analyze. Their result was a much lower runoff value of 0.00053 ft/d under median streamflow conditions. Interestingly, there is a significant variation between low and high streamflow conditions of 0.00012 to 0.002 ft/d. The higher number is indicative of large amounts of additional discharge from shallow soils and sand layers and from drainage tiles; the lower number is more indicative of discharge from the regional aquifers. The baseflow in both streams was significantly influenced by the withdrawal of municipal water from Lake Bloomington in the Mackinaw River basin and its subsequent discharge into Sugar Creek.

Richards and Visocky (1982) calculated the recharge contribution from leakage to the Sankoty-Mahomet Sand aquifer in the vicinity of the Normal well field to be about 0.0004 ft/d. Their study used a flow-net analysis in which pumpage at the well field was balanced by leakage from overlying aquifers over the area of diversion of ground-water flow. This estimate is highly dependent on a very accurate potentiometric surface map, which is difficult to draw for a small portion of a highly transmissive regional aquifer because of the changes in the potentiometric surface caused by pumpage. A higher recharge rate of 0.001 ft/d also was computed by assuming all the recharge to the well field was coming from the shallower aquifers at which the confining unit was absent. These values likely overestimate the actual leakage contribution because they do not take into account the contribution from regional flow in the Sankoty-Mahomet Sand aquifer, and because the upper sands are more extensive than previously mapped.

An estimate for the leakage into the Sankoty-Mahomet Sand aquifer from overlying aquifers also can be made by determining the mixing ratios of ground water from its two main recharge sources, water entering the Sankoty-Mahomet Sand aquifer from the east and water infiltrating downward from above. Each source has a distinct chemical signature. Ground water entering the model area from the Sankoty-Mahomet Sand aquifer has a chloride content that is roughly ten times greater than that of the water in the overlying Glasford Formation aquifers due to discharge of ground water from more saline aquifers in DeWitt and Piatt Counties (Panno et al., 1994). The mixing of ground water from each source produces a composite chloride value in the downstream portion of the aquifer. Herzog et al. (1995a) estimated the amount of leakage through the aquitard directly above the Sankoty-Mahomet Sand aquifer from the upper aquifers to be 0.001 ft/d. This estimate was based on the assumption that there was a continuous layer of thick diamicton on top of the Sankoty-Mahomet Sand aquifer over the entire area. Because the new geologic maps constructed for this study show several areas in which overlying sands are directly connected to the Sankoty-Mahomet Sand aquifer, it is now clear that the recharge to the Sankoty-Mahomet Sand aquifer is not uniform across the study area. Therefore, the 0.001 ft/d estimate may be too high as an average rate of recharge through the aquitard.

The uncertainty in these recharge estimates makes them unsuitable for direct entry into the ground-water flow model. However, the estimates were useful starting points. Other methodologies, such as a flow-net analysis, were not used because available information was insufficient for these techniques. Because of the types of data available for this study, a sound approach, and the one affording the most confidence in the outcome, was to generate recharge and discharge values during the model calibration itself. The estimated best-fit values from this study are discussed later with the calibration results of the ground-water flow model.

Aquifer/Stream Interactions

Another approach for examining baseflow is to look at the stream-flow data collected from the gaging stations during the 7-day, 10-year low flow (Singh et al., 1988) when the near surface sources of stream-water inflow have dried up and most of the water in the stream is from regional ground-water discharge. Using these

data does not have the uncertainty associated with the hydrograph separation technique. The gages along Sugar Creek show that there is no net gain in low flow from Bloomington to Hartsburg; both have flow values of 15.7 cubic feet per second (cfs). Based on observed downstream flow losses from large wastewater treatment discharges at other locations, Singh et al. (1988) estimated that the flow in Sugar Creek drops to 12.0 cfs between the stations. In addition, a gaging station on Kickapoo Creek, for a similar size drainage basin as that between the two stations on Sugar Creek, shows a low-flow gain of 3.2 cfs. Sugar Creek did not gain flow between Bloomington and Hartsburg, even though small tributary streams entered Sugar Creek along this reach. All of these data, taken together, suggest that Sugar Creek is losing a net minimum of 3 to 4 cfs to the regional ground-water system during low-flow conditions. During mean flow conditions when the mean flow at Hartsburg is 80 cfs, the loss of water may be greater because of higher water levels in the stream and a potentially greater downward gradient to the upper aquifer. Based on the increase in drainage basin size, Singh et al. (1988) estimated that Sugar Creek gains an additional 5.7 cfs of low-flow before joining with Salt Creek, just west of the model area boundary.

The increase in 7-day, 10-year low flow in the Mackinaw River through the model area is significantly greater than that of Sugar Creek and Kickapoo Creek, even though the drainage basin areas are approximately the same size. At Congerville, in the northern part of the model area, the low flow is 1.2 cfs; at Green Valley, just west of the model area boundary, the low flow is 25.2 cfs (Singh et al., 1988). After subtracting 3 cfs of treated wastewater discharges, the net increase in low flow is 21 cfs (13.5 mgd). The change in low flow from 1.2 cfs to 25.2 cfs between Congerville and Green Valley is very large for this small drainage area. Based on the geologic data and the results of the aquifer test at Mackinaw, most of this gain is estimated to be downstream of the town of Mackinaw, which also is where the Sankoty-Mahomet Sand aquifer becomes connected to the river. Thus, downstream of Mackinaw, the Mackinaw River is a major ground-water discharge location.

Ground-Water Use

One purpose of the ground-water flow model is to simulate the effects of pumping, and for this accurate data about the amount of ground water withdrawn by existing wells were required. Ground-water use data for municipalities and self-supplied industries are collected annually and stored in the Illinois Water Inventory Program (IWIP) database at the ISWS. Municipal and industrial withdrawals for 1994 were provided by this database for use in the ground-water flow model. Small ground-water users (e.g., residences, farms, and small businesses), which use a comparatively insignificant amount of ground water, were not included in the ground-water flow model. The model area includes several municipalities and industrial users that were not included in the characterization report (Herzog et al., 1995a and b).

In the model area, 82 municipal, subdivision, and industrial wells at 33 facilities had pumpage more than 2,000 gallons per day (gpd) in 1994; 2,000 gpd was selected as the minimum amount of pumpage to include in the model after review of the IWIP data. Though 2,000 gpd is a fairly small amount (1.4 gpm), these data were added to the model because the data were available. Pumpage from domestic wells is widely scattered in low quantities and was accounted for in the model by adjusting the amount of infiltration from precipitation. Figure 22 presents the locations of the pumping wells used in the ground-water flow model and a graphic description of the magnitude of their 1994 withdrawals. A summary of the 1994 pumpage data used in the ground-water flow model is appendix E. One center-pivot irrigation system, just west of McLean, was in use during 1994 in the model area. Approximately 240 acres were irrigated by this system, and the annual ground-water withdrawn from the irrigation well was estimated to be 0.2 mgd.

According to William Simmons, Associate Professor of Soil and Water Management at the University of Illinois (personal communication, 1997), irrigation expansion seems to have ceased in the study area. The largest increase in the amount of irrigation occurred as a result of the last two droughts, the most recent and severe being

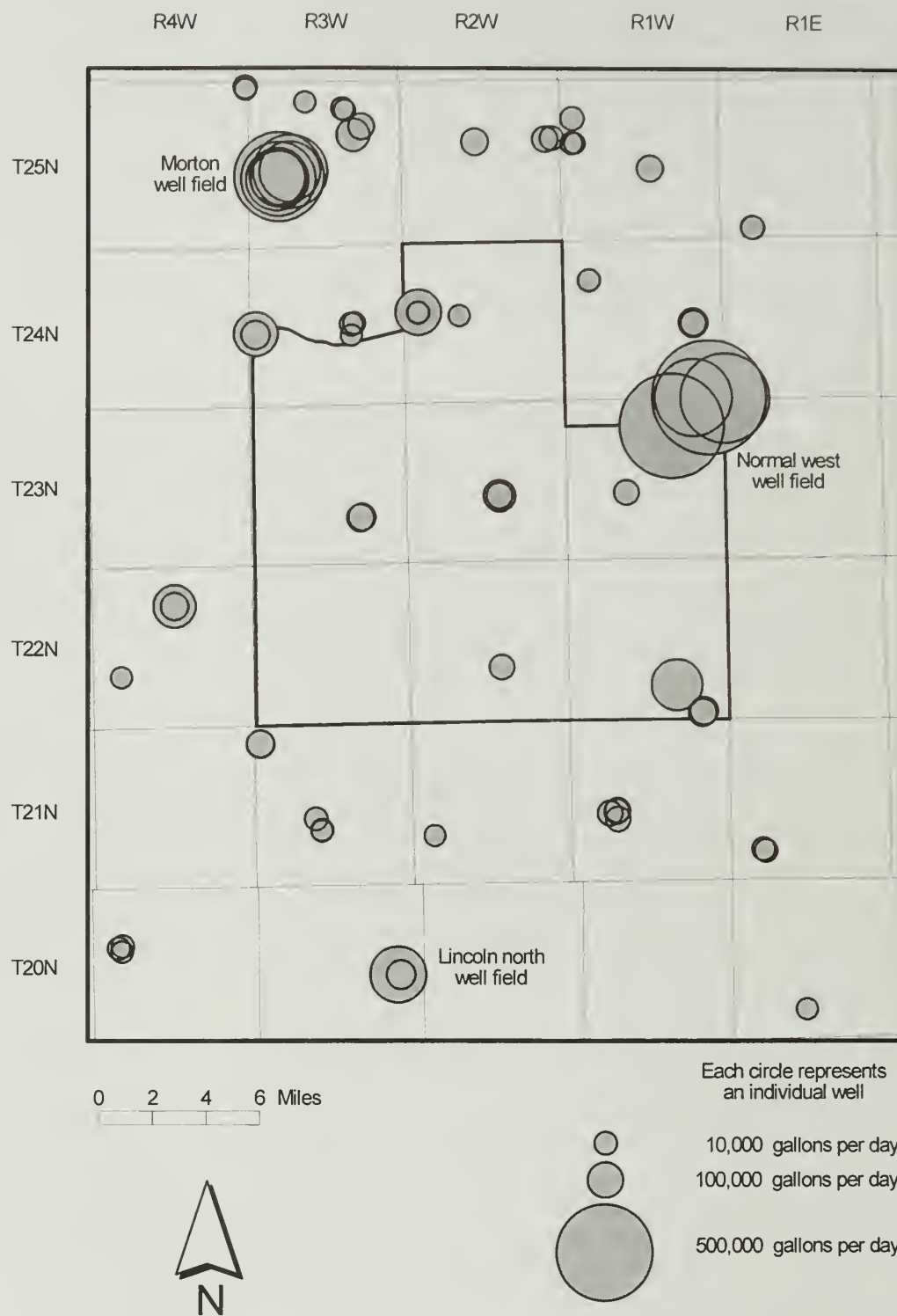


Figure 22. Ground-water withdrawals (gpd) in the model area in 1994

in 1988. Since then, expansion of irrigation into areas with heavy soils, such as the study area, have been minimal. Irrigation has expanded in Illinois during the last 15 years because of increases in seed production by farmers. Most seed companies that contract these services to individual farmers require an irrigation system in place to prevent loss of the crop. Because this market is now near saturation, little expansion is expected in the future. In 1997, a new center-pivot system was installed in the study area, bringing the total number of irrigation systems in the study area to two.

Updated Conceptual Understanding of the Ground-Water Flow System

The collection and analysis of data for this study have permitted the development of an updated understanding of the ground-water flow system. In the model area, ground water in the Sankoty-Mahomet Sand aquifer generally flows from the southeast, where the Mahomet and Kenny Bedrock Valleys enter the study area, toward the Illinois River either westward through the Havana lowlands, or northward through the Mackinaw Bedrock Valley. The hydraulic gradient east of the study area is extremely flat, which indicates a very low leakage rate through the overlying aquitard and/or a very high transmissivity. However, as the flow enters the study area and the aquifer volume increases, the gradient steepens, indicating an increase in the amount of recharge entering the aquifer or a decrease in transmissivity. All of the evidence points toward an increased recharge component. New hydrogeologic mapping shows the easternmost connections of the Sankoty-Mahomet Sand aquifer with the two upper aquifer layers to be in T22N, R2W (figure 19). These connections with the upper aquifers correspond to the area at which increased gradients appear in the potentiometric surface of the Sankoty-Mahomet Sand aquifer as compared to relatively flat gradients in the area east of the model area (figure 16). This is the same area in which the upper reach of Sugar Creek is losing a significant amount of water and is directly connected to the Glasford aquifer. Also at this location, the chloride concentration decreases by one order of magnitude (figure 23), and the concentration of modern carbon isotopes increases in the Sankoty-Mahomet Sand aquifer (Holm, 1995; Herzog et al., 1995a). All of these data indicate there is a significant influx of low-chloride, modern water from the upper aquifers into the Sankoty-Mahomet Sand aquifer.

In the western and northern parts of the model area, the connections with shallower aquifers become more numerous (figure 19). At most of these locations there also are changes in water level, chloride concentration, and modern carbon isotope concentrations in the Sankoty-Mahomet Sand aquifer that indicate a vertical connection with downward leakage. Along the ground-water divide (figure 20), there are no connections with the shallow aquifers and the older, higher chloride water from east of the model area has penetrated into the model area as far north and west as the area just north of Minier (Herzog et al., 1995a). Down-gradient of the connections, however, the chloride concentration remains low, which suggests that the influx through these windows provides the majority of the recharge in these areas. Water from the Normal well field, which is located at one of these upper aquifer connections, has low chloride values, indicative of water coming from the upper sands.

The first natural ground-water discharge areas encountered by ground-water flow in the Sankoty-Mahomet Sand aquifer are the Mackinaw River in the west central part of the model area and the lower reach of Sugar Creek in the southwestern part of the model area. In areas very close to the rivers, ground water may be flowing upward from the Sankoty-Mahomet Sand aquifer through the upper aquifers before reaching the streams. Flow in the Sankoty-Mahomet Sand aquifer continues across the western model boundary to Mason County and ultimately to the Illinois River between Havana and Pekin. The bedrock high between Pekin and Tremont forces ground water to flow around it, so the ground water in the eastern portion of the model area flows north toward the Morton well field and ultimately to the Illinois River at East Peoria. The Mackinaw River is isolated from the Sankoty-Mahomet Sand aquifer in the northern and eastern portions of the model area by thick aquitard layers, so at these locations the Mackinaw River's influence on the flow field in the aquifer is limited.

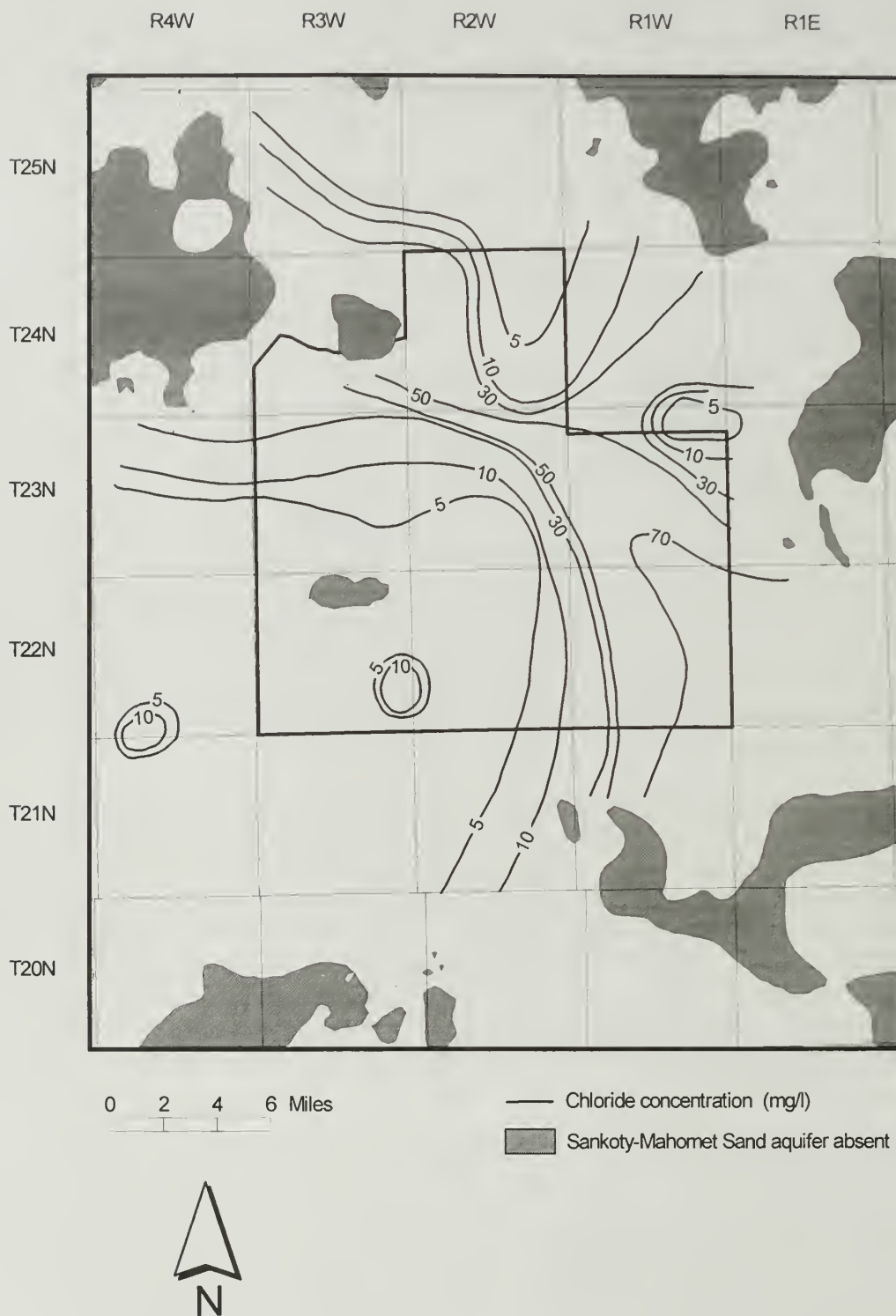


Figure 23. Chloride concentration of the Sankoty-Mahomet Sand aquifer in the study area

The recharge and discharge components will change as different stresses are applied to the system, such as additional pumpage, drought, or flooding along the major rivers. Recharge from infiltration and vertical leakage can, in some cases, be significantly increased by increasing the vertical head gradients through ground-water pumpage. In the Sankoty-Mahomet Sand aquifer, the closer any new production wells are to the areas of direct connection with the upper aquifers, the greater will be the amounts of induced downward leakage from the upper aquifers. Pumpage from a high capacity well next to a gaining stream could cause a reversal of flow by lowering the hydraulic head in the aquifer to a level below the level of the stream. The water induced into the aquifer by this reversal would offset the water removed by withdrawal by the well, but also would reduce the flow in the streams. The locations in the model area at which this could potentially occur include reaches along both the Mackinaw River and Sugar Creek.

Lateral inflow from the east also might be increased into the model area if the drawdown from a new pumping center intersects the eastern boundary of the model area. From a practical standpoint, however, because the Sankoty-Mahomet Sand aquifer is highly transmissive, it would require significant pumpage to noticeably increase flow gradients. In other words, drawdown (and, therefore, the gradient) is minimized due to the high aquifer transmissivity. However, the extent of the cone of depression will be greater than for an aquifer with a lower transmissivity. Analysis of available data suggests that it is likely the Sankoty-Mahomet Sand aquifer operates under relatively nonleaky (relatively little vertical recharge from shallower layers) artesian flow conditions in the southeastern portion of the model area as well as farther east in the Mahomet aquifer. Nonleaky conditions reduce pumping capacity compared to leaky conditions because there is less vertical contribution of water to balance with pumpage and offset the drawdown.

The most reasonable means to maximize production capacity while minimizing impacts to existing wells is to locate pumping centers in areas of (1) maximum aquifer thickness (to maximize transmissivity), (2) maximum available drawdown, or (3) maximum available recharge either when the aquifer is unconfined or when the aquifer is proximal to recharge boundaries.

MATHEMATICAL GROUND-WATER FLOW MODEL

The purpose of a computer-based, mathematical ground-water flow model is to simulate the ground-water flow system in the modeled area. By developing such a model, simulations can be made of what will likely happen to water levels in the model area when the system is modified from its current conditions. In this case, the changes in conditions include adding hypothetical pumping scenarios to the model to simulate the effects of a 15-mgd well field. In complex hydrogeologic situations, such as those found in the model area, a properly calibrated ground-water flow model is the best tool available for making predictions of the effects of added pumpage on the aquifer system and to water levels in existing wells. Development of this type of model incorporates knowledge of the geology, hydrology, and hydraulic properties into the conceptualization of the ground-water flow system.

General Procedure

A model of the ground-water flow system was constructed for the model area using the program Visual Modflow (Waterloo Hydrogeologic, Inc., Waterloo, Ontario, Canada). Visual Modflow is an enhanced version of Modflow (McDonald and Harbaugh, 1988) that includes added features for adding data and presenting results. Modflow, developed by the USGS, is widely accepted and used in the ground-water industry. Modflow is a three-dimensional model that divides the subsurface into a grid of discrete cells, each of which is assigned values for the various hydraulic properties. Wells, rivers, and other boundaries were simulated by adding specified head or discharge conditions to a cell or group of cells. The program calculates a hydraulic head value at each cell using a finite-difference technique for the partial differential equations that combine Darcy's Law with a mass balance expression.

Creation of the many input files necessary to run the model was greatly simplified by the use of the Visual Modflow software package, which has a graphical interface. During the course of this study, faster computers and extended memory versions of Visual Modflow became available that allowed for a much larger data set to be used than previous PC versions and reduced the model run times. The software programs Surfer (Golden Software, Golden, CO), Canvas (Deneba Software, Miami, FL), and ArcView (ESRI, Redlands, CA) were used to prepare the final figures from the model output.

Model Design

The model grid (figure 24) extends over the model area, which was expanded beyond the original study area to include more of the natural aquifer boundaries and more of the area influenced by a potential large-capacity well field. The grid covers an area 30.5 miles east to west by 36.5 miles north to south and is divided into 122 columns, 146 rows, and 7 layers for a total of 124,684 discrete mathematical cells. Each model layer generally corresponds to an aquifer or aquitard layer. The horizontal dimensions of each square cell are a constant 1,320 ft x 1,320 ft (40 acres) with the vertical cell thicknesses varying from 2 feet to more than 100 feet, depending on the hydrogeology.

The elevations for the tops and bottoms of each of the aquifer units were input into Visual Modflow directly from EarthVision output files using the same grid dimensions. This approach greatly improved the ability to accurately introduce hydrogeologic data into the flow model. Using EarthVision eliminated three intermediary data manipulation steps of digitizing hand-drawn contours, gridding digitized lines, and manually editing the grid files in problem areas. By cutting off intersecting trends in the hydrogeological surfaces according to a set of erosional and depositional rules and maintaining a minimum thickness when a layer was absent, the use of EarthVision also eliminated the potential problem of having negative thicknesses in the model. The land surface elevation was input into the model from a gridded surface created from the digitized contour lines on the USGS 7.5-minute topographic maps. The land surface grid had to be hand edited along the simulated streams to accurately represent their elevations.

The model was run under steady-state, or equilibrium, conditions. The model stopped running when the computed water levels changed by less than by a set criterion. The criterion chosen for this model was 0.01 feet because it is low compared to the overall 100-foot head change across the model and because it produces a very small discrepancy between the volume of water flowing in and out of the model. There is no way to know beforehand what the computed water levels in the model will be; therefore, the model has to work from some arbitrary water-level distribution as an initial condition. For this model, a constant value of 700 feet was used as the starting head because it is close to the expected water levels and keeps all of the model layers initially saturated. The drought scenario, discussed later, was run under transient conditions using the water levels from the steady-state model as its initial condition.

Model Layers

Each hydrogeologic layer was assigned to a model layer (figure 25). For computational purposes, each model layer must be continuous across the model area. If a hydrogeologic layer does not exist over a portion of the layer, the corresponding cells in that model layer were assigned a minimum thickness of 2 feet and were given the hydraulic properties of the next existing layer above. The major exception to this convention occurred in the river valleys in which the properties of the next existing layer below was used because the upper layers have been eroded. The relationships between the hydrogeologic layers as represented in the model are shown on figures 26 and 27, which are representative cross sections through the center of the model, and on figures 28-33, which are individual maps of each model layer.

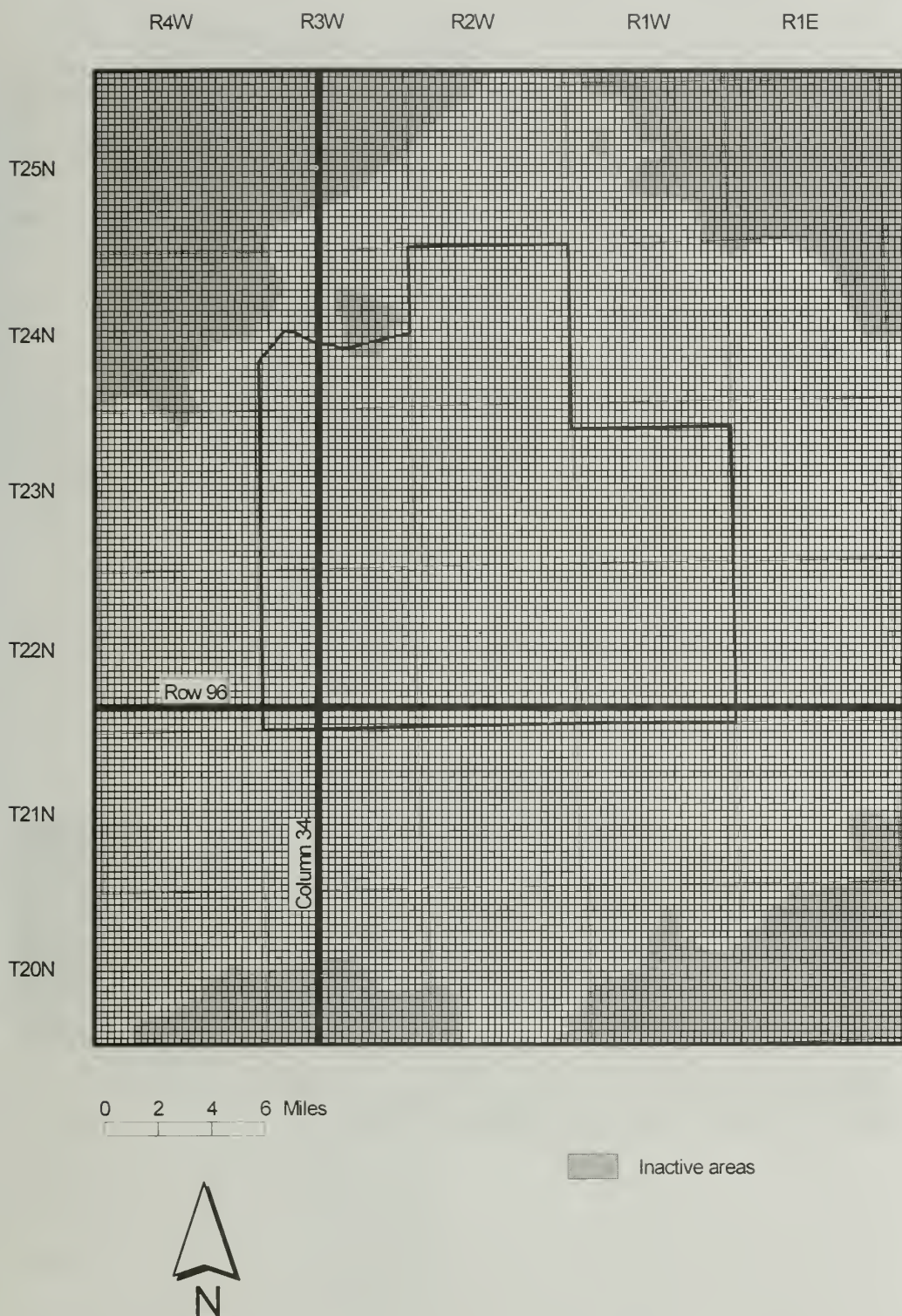
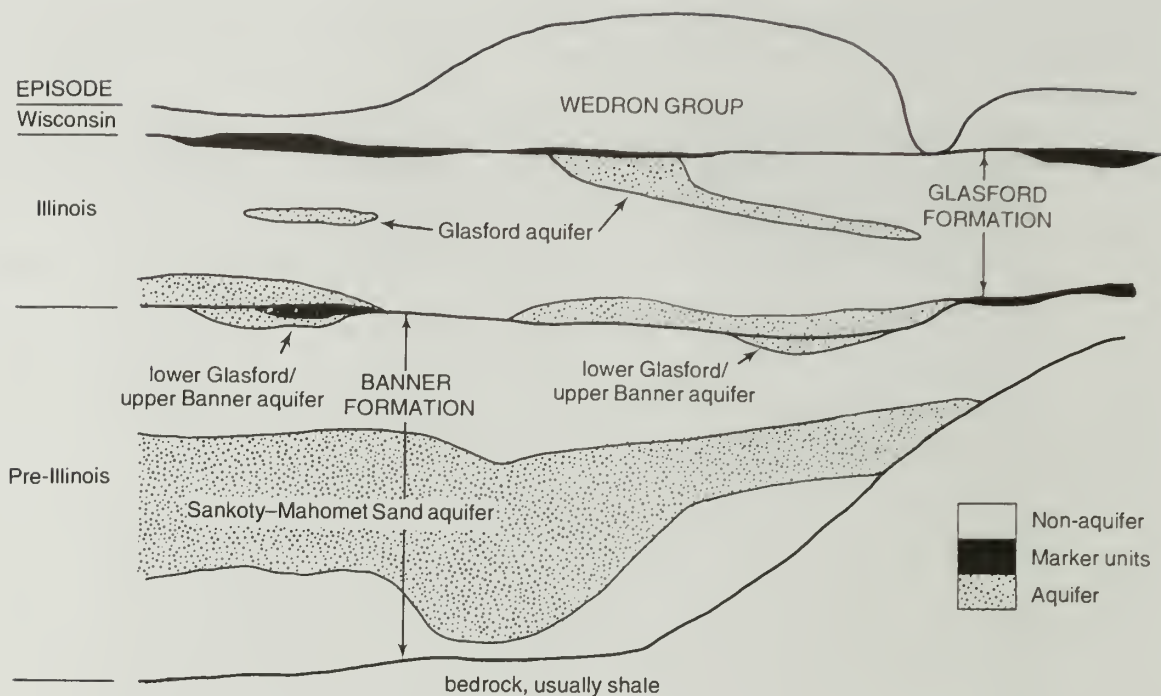
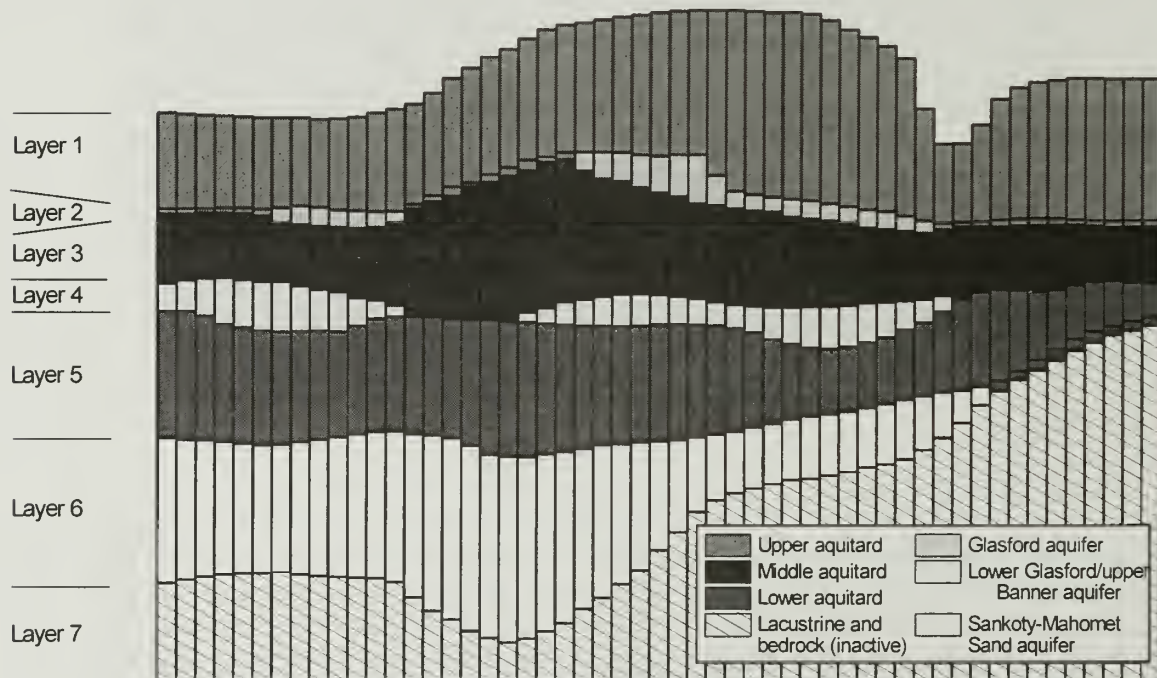


Figure 24. Grid used in the ground-water flow model and location of model cross sections



a. Simplification used in the hydrogeologic representation



b. Representation of the hydrogeologic layers by the model

Figure 25



Figure 26. East-west cross section through model row 96 (see figure 24)

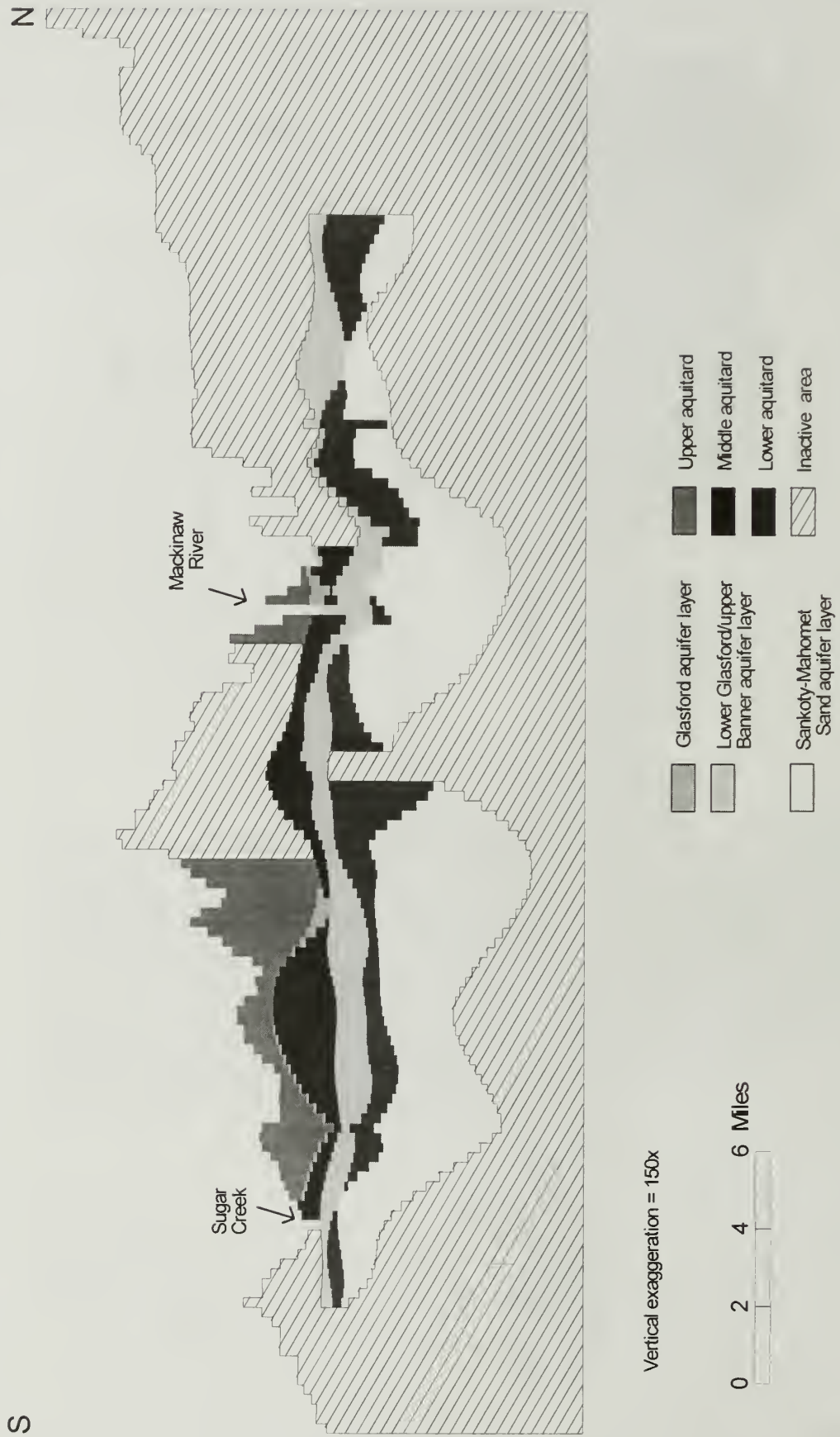


Figure 27. North-south cross section through model column 34 (see figure 24)

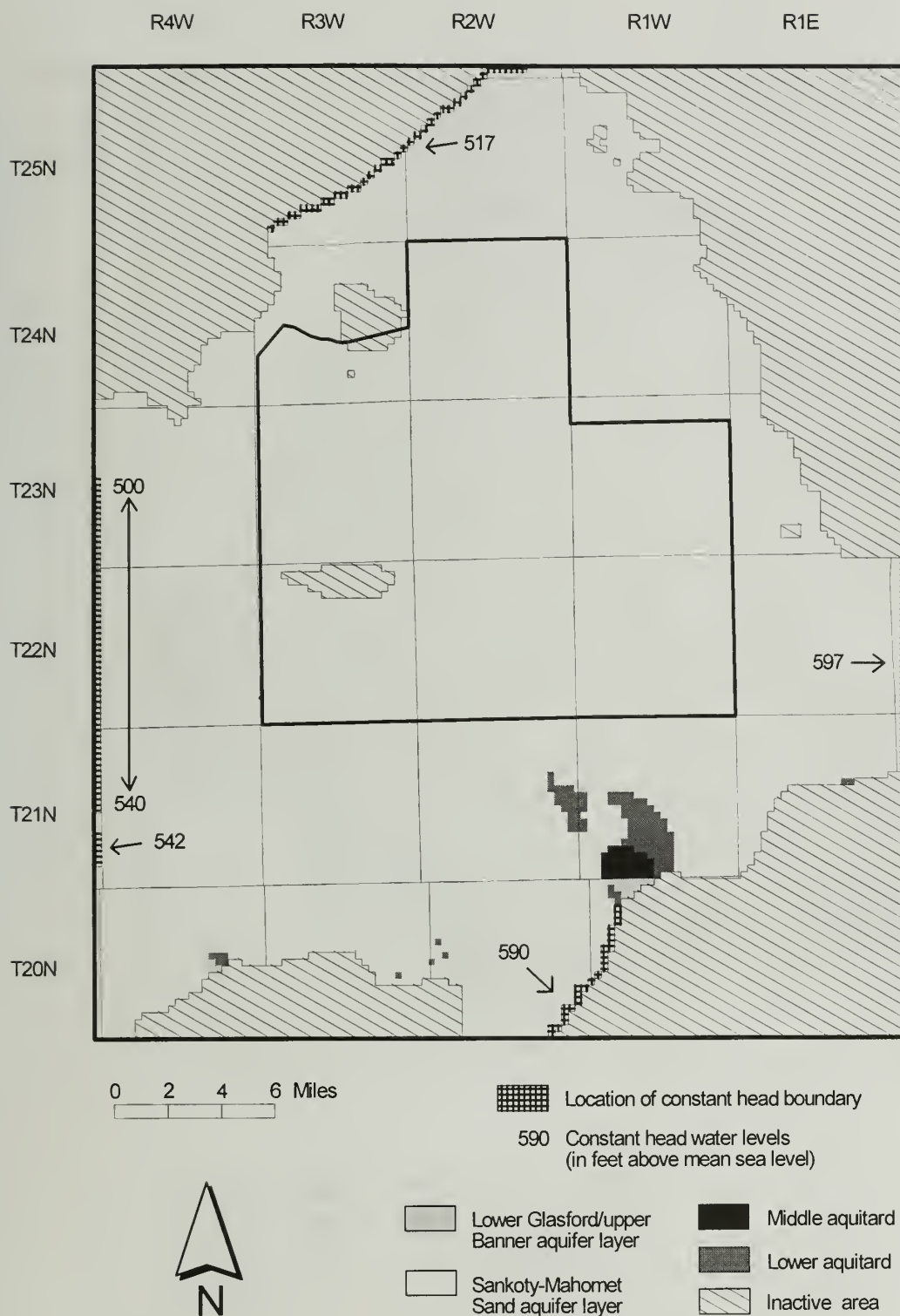


Figure 28. Hydrogeologic layers in layer 6 of the ground-water flow model

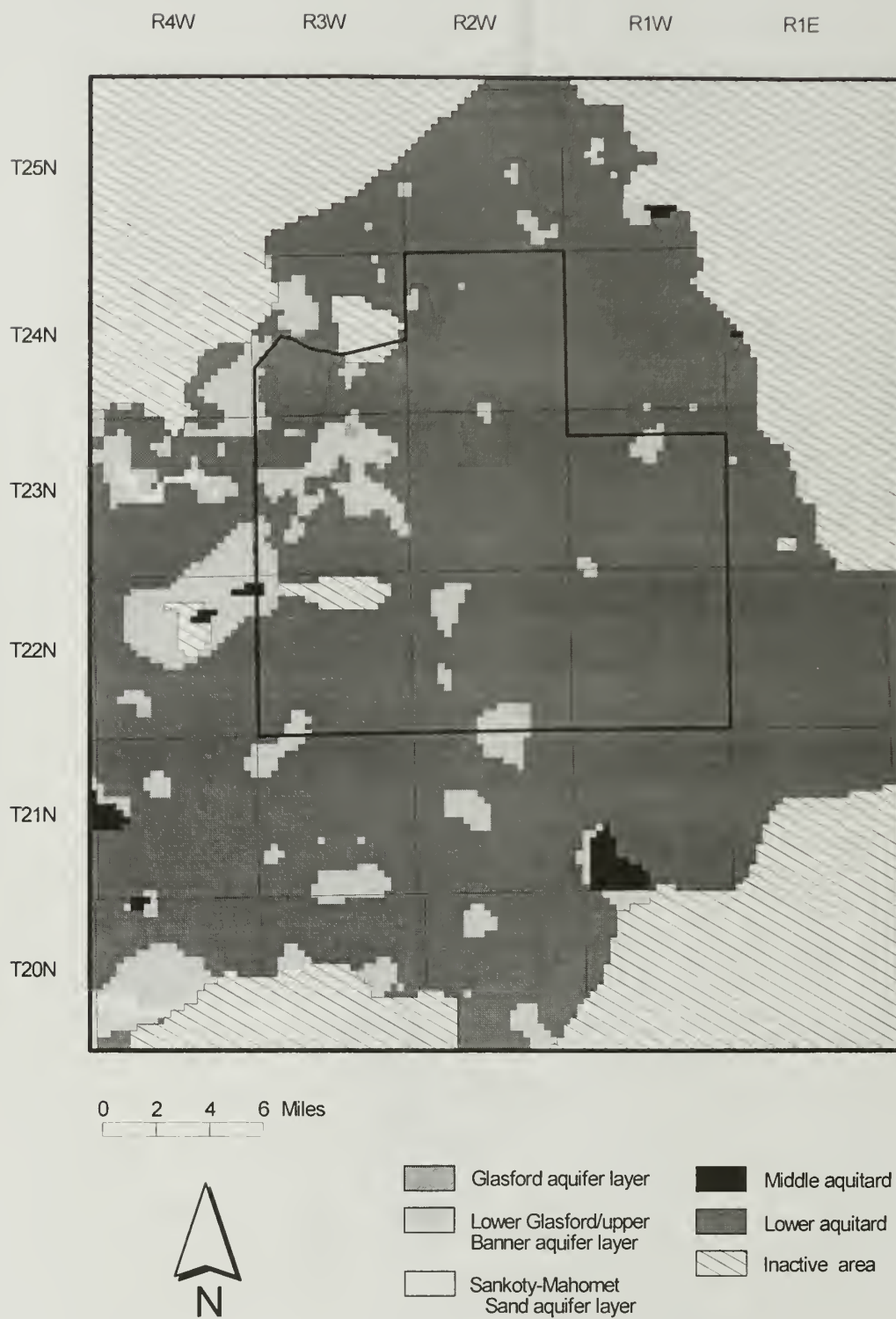


Figure 29. Hydrogeologic layers in layer 5 of the ground-water flow model

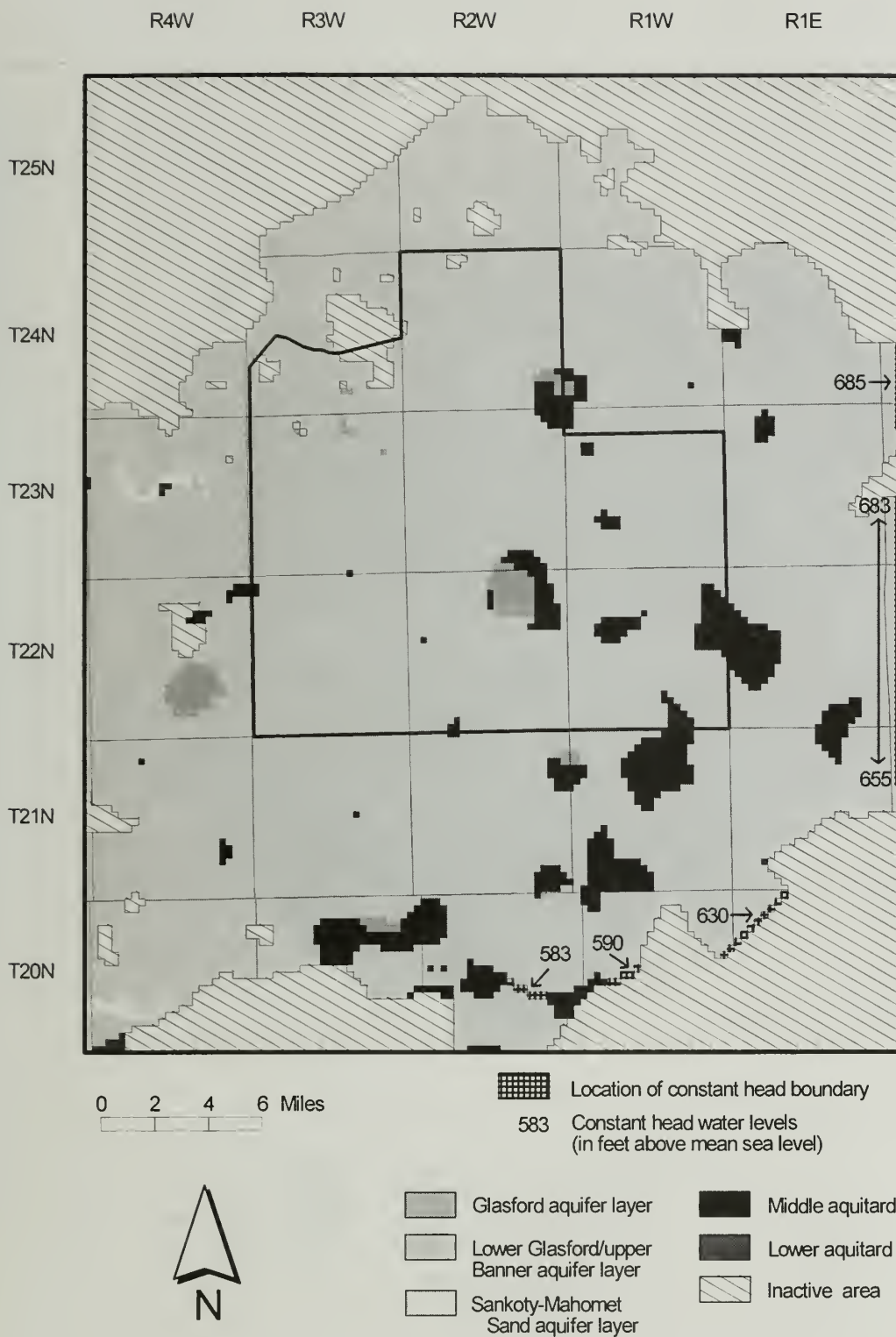


Figure 30. Hydrogeologic layers in layer 4 of the ground-water flow model

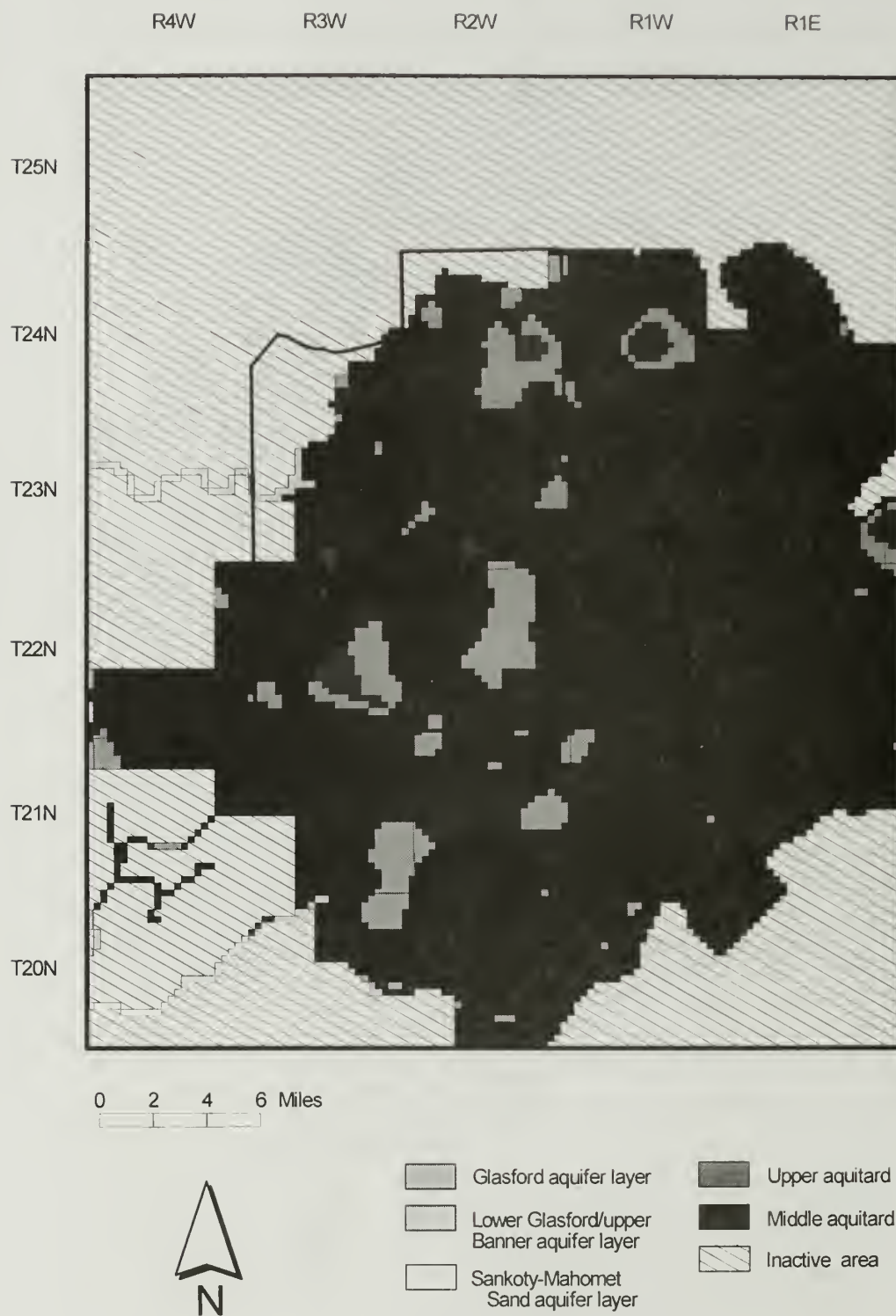


Figure 31. Hydrogeologic layers in layer 3 of the ground-water flow model

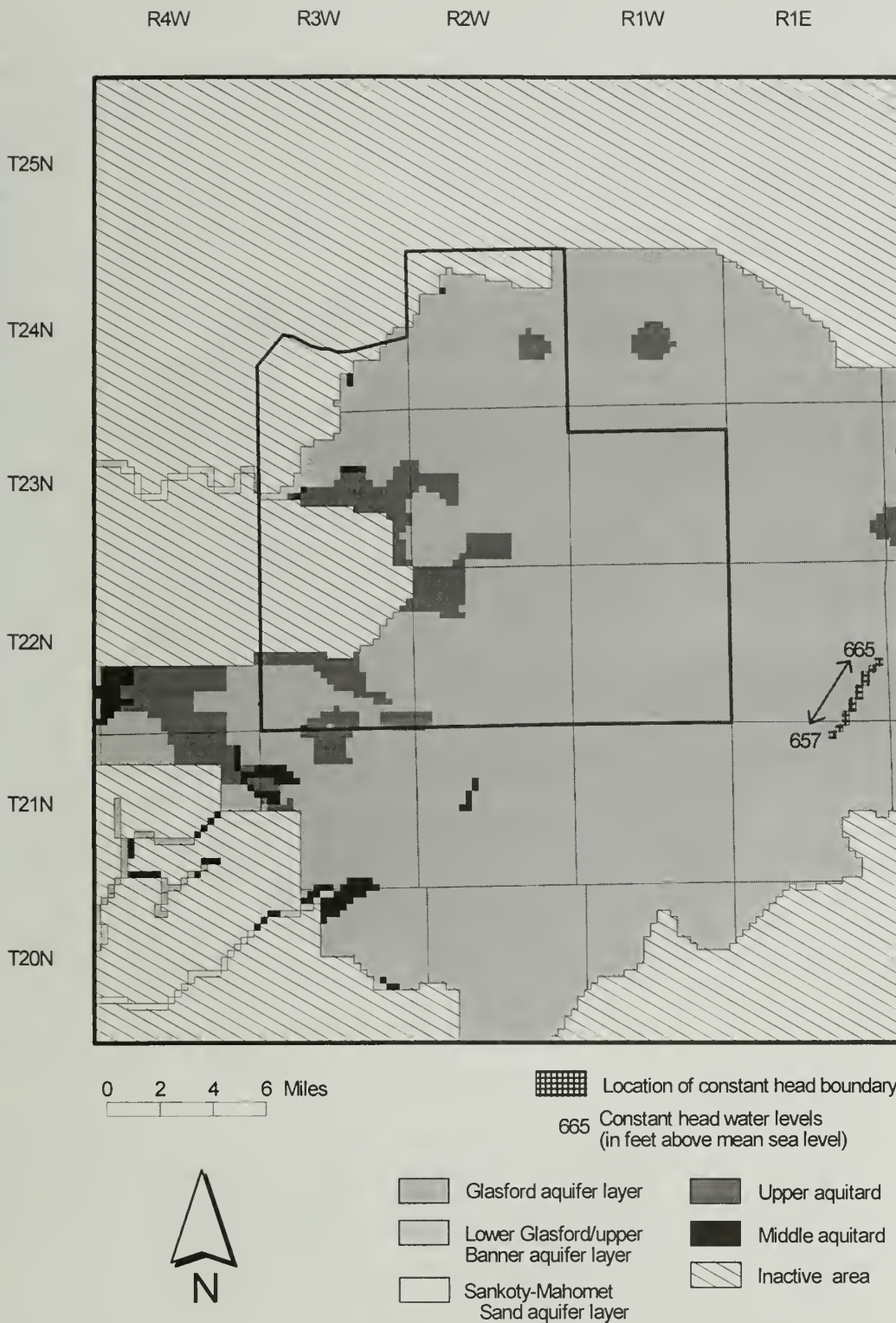


Figure 32. Hydrogeologic layers in layer 2 of the ground-water flow model

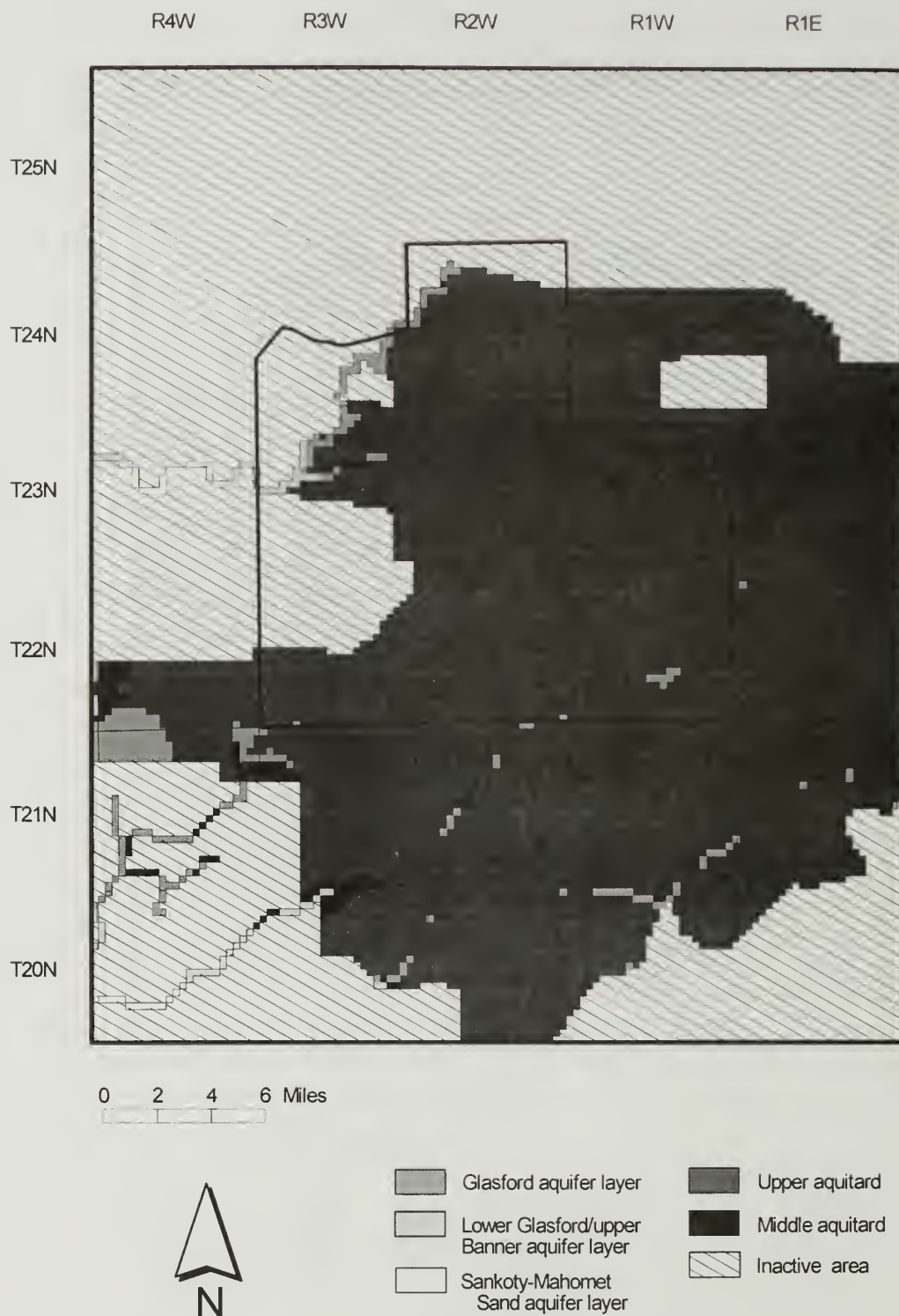


Figure 33. Hydrogeologic layers in layer 1 of the ground-water flow model

The Sankoty-Mahomet Sand aquifer layer is represented by layer 6 of the model (figure 28) and in the upper layers along the Mackinaw River and Sugar Creek in R4W where the Sankoty-Mahomet Sand crops out. The inactive area shown on figure 28 is where the aquifer does not exist beyond the valley walls or it is behind a boundary condition in the model that simulates regional flow into or out of the model area. The lower Glasford/upper Banner aquifer layer is represented by layer 4 (figure 30) and extends farther to the east than the Sankoty-Mahomet Sand aquifer layer because it laps onto the bedrock surface at a higher elevation or continues beyond the bedrock valley walls. The Glasford aquifer layer is represented by model layer 2 (figure 32). Large portions of this aquifer layer and of the aquitard layers above and below it on the west and north sides of the model, except along the river beds, were made inactive where the layers are naturally dry. During the calibration process, much of the area in model layer 2 where the Glasford aquifer is absent was assigned a hydraulic conductivity between that of the aquifer and that of the overlying aquitard. Because these zones are only 2 feet thick, they do not appear on the cross sections.

Model layers 5, 3, and 1 (figures 29, 31, and 33) represent the aquitard layers. With advances in computer speed and memory and the recent ability of various Modflow preprocessor and postprocessor software programs to display vertical slices accurately, it was decided to model the aquitard layers explicitly instead of simply assuming leakage values for them. Having the aquitard layers as model layers gives better control over the hydraulic properties during input and allows for horizontal flow within the aquitards. An important feature of these model layers are the areas in which the aquitard is absent and the aquifers are directly connected. The uppermost surface shown in the cross sections (figures 26 and 27) represents the land surface. Although the land surface elevation is not used directly in any of the calculations, having it in the model allows the user to see if the computed water table is too high or too low.

The bottom, or seventh, layer of the model represents all of the material below the Sankoty-Mahomet Sand aquifer layer down to an elevation of 300 feet, which is just below the lowest elevation in the Sankoty-Mahomet Sand aquifer layer of 301 feet. This model layer is inactive in the model, but it is necessary to preserve the integrity of the cross-sectional views in Visual Modflow. The bedrock largely consists of Pennsylvanian shales, which, with the lacustrine material between the lowermost sand and the bedrock, have a very low permeability and would not contribute or convey a significant amount of ground water in the modeled system. Thus, the top of model layer 7 is a no-flow boundary. However, if the chemistry of ground water from the bedrock is significantly different from that of the overlying aquifer, this seemingly insignificant amount of flow could have a noticeable impact on the water quality of the Sankoty-Mahomet Sand aquifer. East of the model area in Piatt County, Panno et al. (1994) attributes an increase in chloride content in the Mahomet Sand aquifer to discharge of saline water from the bedrock.

Initial Aquifer Property Values

The final hydraulic conductivity values used in the model were arrived at during the calibration process, whereby an initial set of values were adjusted to minimize the amount of error between simulated and measured head values and flow budgets. Because the hydraulic property distributions for each model layer could not be imported directly from EarthVision into Visual Modflow, all hydraulic property reassignments were done manually cell by cell. An initial hydraulic conductivity value of 335 ft/d for the Sankoty-Mahomet Sand aquifer layer was input into the model based on the similar results from the aquifer tests at Emden (312 ft/d) and Mackinaw (355 ft/d) conducted as part of this study and from tests at the Normal well field, which had a mean of 319 ft/d (Herzog et al., 1995a and b). The variation in the results from the aquifer tests at Normal wells—100 (389 ft/d), 102 (351 ft/d), and 103 (216 ft/d)—indicate that there is some variability within the aquifer, even over relatively short distances. For the Glasford and the lower Glasford/upper Banner aquifers, an initial hydraulic conductivity value of 170 ft/d was input into the model based on other aquifer tests in the region (Herzog et al., 1995b; Kohlhase, 1989).

Data for the hydraulic conductivity of the aquitards were not available. The aquifer test near Emden did not show any leakage through the aquitard, whereas the test at Mackinaw showed significant leakage (Herzog et al., 1995a). The vertical conductivity of 0.03 ft/d for the aquitard at the Mackinaw site may be higher due to the aquitard being thin and possibly fractured, and to leakage directly from the upper aquifer in nearby areas in which the aquitard is absent (figure 29). The lack of leakage during the Emden test and large head differences (up to 80 feet across the lower aquitard) in T21N, R1E suggest that the vertical hydraulic conductivity of the aquitard is extremely low. An initial value would be on the order of 0.0001 to 0.001 ft/d, which is in the middle of the range for glacial till suggested by Freeze and Cherry (1979). Marsh (1995) used a value of 0.0005 ft/d for a model in the eastern Mahomet Valley aquifer based on a leakage value suggested by Visocky and Schicht (1969).

Recharge and Discharge Boundary Conditions

Model Boundaries

To reduce the amount of error and uncertainty in a model, it is ideal to extend the boundaries of a model to the natural boundaries of the aquifer. However, as shown in figure 16, the original study area encompasses only a small portion of the Sankoty-Mahomet Sand aquifer system. By adding a border 6 to 12 miles wide around the study area, the potential error was reduced by taking advantage of the buried bedrock valley walls to the northwest, northeast, and south (figure 4) where the Sankoty-Mahomet Sand aquifer layer pinches out. These bedrock valley walls are assumed to be no-flow boundaries and were inactive. Inactive areas are not used in any of the numerical calculations. The cells beyond the bedrock valley walls also were made inactive. As discussed under Model Calibration, the Sankoty-Mahomet Sand aquifer within the Danvers Bedrock Valley was made inactive because it does not appear to be contributing a significant amount of water to the main body of the aquifer.

The remaining perimeter on the southeast, north, and west sides of the model area are open to flow into and out of the regional aquifer system (figure 16). This flow was simulated in the model with constant-head boundaries, which take out or add to the model as much water as is necessary to maintain the prescribed hydraulic head at the boundary. The steeper the gradient is to a constant-head cell, the greater the volume of water that will be removed or added by the boundary cell. Boundary errors can occur if a production well is placed too close to a constant-head cell because additional water will be added to the model without creating the proper amount of drawdown.

Water-level values for the constant-head cells along the model edges were selected from the potentiometric surface map (figure 20). Constant heads were assigned to the cells along the 597-foot equipotential in the Mahomet Bedrock Valley on the east side, the 590-foot equipotential in the Kenny Bedrock Valley on the southeast side, and the 517-foot equipotential in the Mackinaw Bedrock Valley on the north side (figure 28). The areas beyond the constant-head cells in the Kenny and Mackinaw Bedrock Valleys were made inactive. The inactive area in the Mackinaw Bedrock Valley encompasses the city of Morton's well field, which could not be simulated accurately without extending the model area farther north. Along the central part of the western boundary, flow is mainly to the northwest, cutting across the model edge and precluding the use of a single equipotential as a boundary. Therefore, a line of constant-head cells was placed along the western edge, with head values ranging from 500 feet near the Mackinaw River to 542 feet near Prairie Creek. Flow out of the southwest corner of the model area is dominated by Sugar and Prairie Creeks so constant-head boundaries were not needed.

Constant-head boundary conditions of between 630 and 680 feet were placed along parts of the east and southeast edges of the lower Glasford/upper Banner aquifer layer in which water is flowing into the model area, based on the contours in figure 19. A short line of constant-head boundary conditions also were placed in the Glasford aquifer layer at the southeast corner of the model at which water is flowing in at elevations between

657.5 and 665 feet (figure 18). Specified head boundary conditions for these two aquifer layers are not needed in the northern and western edges of the model because they are absent, dry, or directly connected to the Sankoty-Mahomet Sand aquifer layer that is controlling the water levels.

Streams

Three major streams, which are important to the hydraulic budget of the aquifer systems, cross the model area: the Mackinaw River, Sugar Creek, and Kickapoo Creek (figure 1). Because most of the smaller tributaries within the model area interact only with the ground water in local flow systems, only the larger tributary streams with potential significance to the deeper ground-water flow system were modeled. These modeled tributary reaches start upstream of where the streams begin to have a reliable ground-water contribution as defined by the 7-day, 10-year low-flow condition (Singh and Adams, 1973). In the active portion of layer 1, these tributaries include the Little Mackinaw River, Prairie Creek, and the West and Middle Forks of Sugar Creek. The upper portion of the Mackinaw River is inactive because it only interacts with the thick surficial aquitard, which also is inactive in this area .

The rivers were input into model layer 1 (figure 34) using the river package of Visual Modflow, which simulates a head-dependent flux boundary condition. Flow into or out of a river cell is dependent on the river stage, the river bottom elevation, the river bottom conductance, and the hydraulic head within the cell. The river stage data were taken directly from the USGS topographic maps in which the elevation contours cross the streams; these data were used to edit the land-surface elevations. The river bottom elevations were assumed to be 2 feet below the river stages. The conductance of the river bottoms were computed using a width of 50 feet for the larger streams and 20 feet for the smaller streams, and assuming the hydraulic conductivity of the river bottom is the same as the hydrogeologic layer it is passing over. This assumption is based on numerous scour features observed along the river beds. The computed conductances for streams running over the aquifer and aquitard layers were 7,430,000 cubic feet per day (ft^3/d) and 1,000 ft^3/d , respectively. The aquifers crop out at the land surface, as shown on figure 33, at locations where an aquifer layer exists in model layer 1. This occurs exclusively in the river valleys except at an area along the western edge of the model.

Infiltration from Precipitation

Infiltration from precipitation is simulated in the model using the constant flux boundary condition provided by the recharge package of Visual Modflow. The flux from this recharge source is applied to the uppermost active layer in the model. For example, the recharge immediately north of the Mackinaw River at the western edge of the model area (figure 35) is applied directly to the lower Glasford/upper Banner aquifer layer in layer 4 because the layers above are inactive at this location. Recharge values used in other flow models of the Mahomet and Sankoty-Mahomet Sand aquifers range from 0.00017 ft/d (0.75 inches/year) in Vermilion County, where the aquifer is well confined (Marsh, 1995), to 0.0023 ft/d (10 inches/year) in Mason County, where the aquifer is under water-table conditions (Clark, 1994). Because the aquifers are overlain by fine-grained material in almost all of the model area, except along some of the rivers, and because of the previous estimates and the new estimates, an initial recharge value of 0.00023 ft/d (1 inch/year) was used in the model. As part of the calibration process (discussed later), this value was lowered for most of the model area to 0.00018 ft/d (0.78 inches/year). The rate in small, specific areas was either raised or lowered during calibration (figure 35).

Production Wells

The effects of larger production wells in the model area were simulated using the well package of Visual Modflow which is a constant flux boundary condition that removes water from the middle of a cell in the model. Forty production wells pumping a total of 495,026 ft^3/d (3.7 mgd) were simulated in the model. Some of the

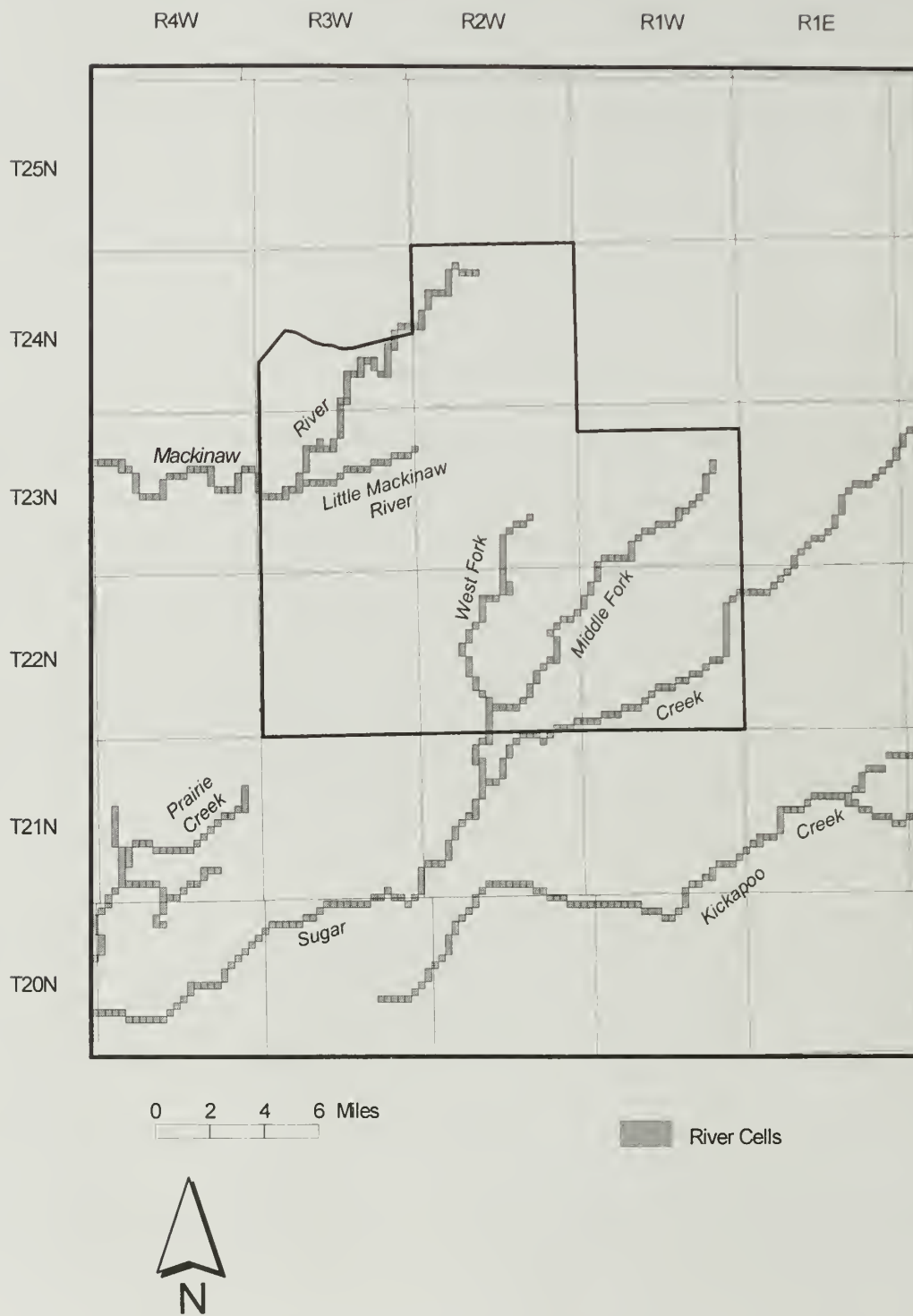


Figure 34. River cells in layer 1

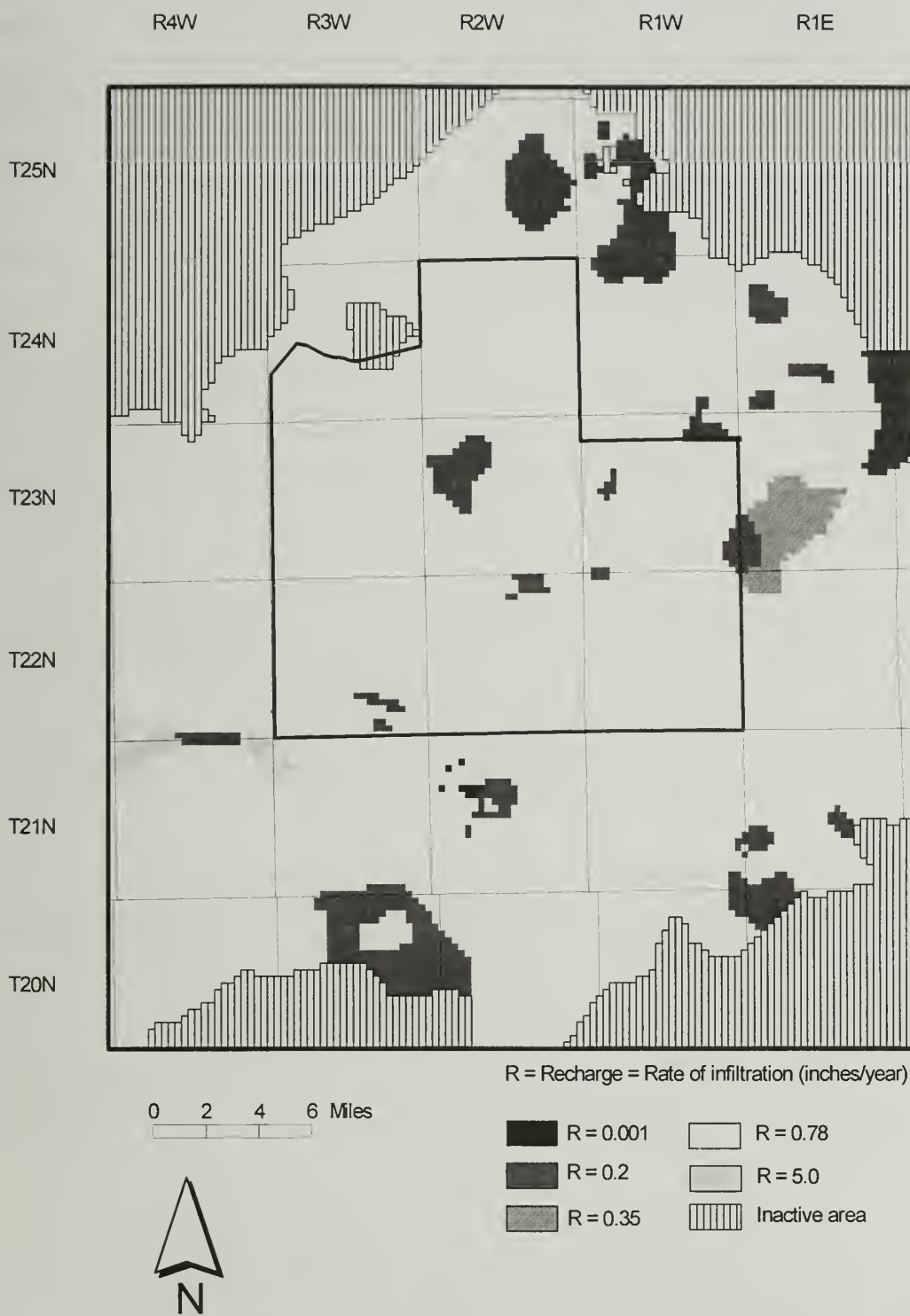


Figure 35. Recharge distribution for the uppermost active layer

wells shown on figure 22 were combined because they were located in the same cell, but others, such as the Morton wells, were in inactive areas of the model. All the production wells withdraw water from only one aquifer layer, except Normal well 100, which withdraws water from both the lower Glasford/upper Banner aquifer layer and the Sankoty-Mahomet Sand aquifer layer. The town of Normal well fields in the Glasford aquifer east of the study area have the potential to affect the study area. However, Kempton (personal communication, 1972) indicates that the sand these wells are completed in may be isolated from the Glasford sands in the model area.

Model Calibration

Model calibration is the process of iteratively running the model while selectively changing model input parameters until a solution is reached that satisfies a prescribed error criterion. The complexity of the hydrogeologic layers and the potentiometric surfaces made this model very difficult to calibrate. Several generations of the model were constructed, with the last generation requiring 312 different runs before the model was calibrated. The calibration process was greatly aided by the updated conceptualization of the hydrogeology with the division of the sands above the Sankoty-Mahomet Sand aquifer layer into the Glasford and the lower Glasford/upper Banner aquifer layers. However, the added geometric complexity made the model more difficult to run because small changes in a parameter value often made the program "crash" for purely mathematical reasons. Parameter value changes often had to be made in different phases over small areas until the problem cells were found.

A prescribed error criterion is not the same for all ground-water flow models. Rather, it depends on the characteristics of the flow system, such as the overall head change across the model, and the complexity of the potentiometric surface. The best way to assess how well a model is calibrated is to compare the computed water levels and flow budgets with measured data from the field. Quantifying the error in water levels can be done either by subtracting the computed water levels from the measured water levels in an observation well network, or by subtracting the computed potentiometric surface from that interpreted from measured water levels. Flow budgets are more difficult to compare because the volume of water moving through an aquifer cannot be directly measured, except where the aquifer discharges or receives water from a stream.

The error associated with collecting and interpreting the field data also has to be considered when setting an error criterion. Water levels from wells can be affected by measurement and surveying errors, or by wells that are not well developed, especially domestic wells. A potentiometric surface map constructed from measured water levels has additional error due to contour smoothing. The smoothing error between the mass measurement water levels in the Sankoty-Mahomet Sand aquifer and the potentiometric surface from them had a mean of -0.64 feet and an absolute mean of 4.76 feet. For just the observation wells drilled as part of this study, the errors were lower with a mean of 0.10 feet and an absolute mean of 2.67 feet. These errors suggest that the Sankoty-Mahomet Sand aquifer layer has a significant amount of hydrogeologic variation at a local scale, so producing a model with a lower error would be unreasonable unless enough additional observation wells were drilled to examine these local scale variations.

Considering the 105-foot drop in the potentiometric surface of the Sankoty-Mahomet Sand aquifer across the model area, the complexity of the hydrogeology, and the uncertainty associated with the potentiometric surface maps created from the field data, a goal was set to produce a model with a mean absolute error at the observation wells of less than 5 feet; there were to be no significant trends or spikes in the distribution of the error. Although a model with a 5-foot error would not match every complexity, it would match the conceptualization of the flow system very well. Some additional errors will result when there are steeper hydraulic gradients because the computed head is averaged over a 40-acre grid cell, instead of being a point value at which the observation well is located.

Approach

The process of calibrating the ground-water flow model started with activating only the Sankoty-Mahomet Sand aquifer layer, which dominates the flow in the model area. After a reasonable match with the general water-level trends and elevations of that layer were achieved as a one-layer model, and the basic boundary conditions were set, the upper layers of the model were activated. The process of fine-tuning the different aquifer parameters could then begin. In addition to constantly checking the configuration and mean absolute error of potentiometric surfaces throughout the process, the relationships of the water table to the land surface and the gains and losses of ground water to the streams were monitored. The process was facilitated by the ability of Visual Modflow to allow a user to place observation wells in the model and automatically compute the error statistics after each run.

Results

Calibration was successfully completed with the reduction of the mean absolute error in water levels to 4.99 feet using the hydraulic conductivity values listed in table 3. A plot of the calculated hydraulic heads of the model versus those observed in the observation wells is shown on figure 36. Four of the underpredicted heads (below the line) are from observation wells in the shallow aquifers in which potentiometric surfaces are not as well defined and isolated sands were combined to make the layer. These shallow water levels also contribute to the mean error being -2.10 feet instead of being closer to zero. The root mean squared error is a relatively small 6.33 feet. The potentiometric surfaces of the three aquifers computed by the model and the associated error distributions with the measured surfaces are shown in figures 37-42. Generally, the error maps show a good match between each set of potentiometric surfaces without any significant trends that would affect the usefulness of this model for its intended purposes.

The most difficult error to reduce in the Sankoty-Mahomet Sand aquifer layer (figure 38) occurs where the gradient more than triples from 0.00022 ft/ft in T22N, R1W to 0.00078 ft/ft in T23N, R2W. This is the location at which the aquifer is undergoing a transition from the well-confined system in the Mahomet Bedrock

Table 3. Hydraulic Conductivity Values Used in the Calibrated Model

<i>Hydrogeologic layer</i>	<i>Principal model layer</i>	<i>Horizontal hydraulic conductivity (ft/d)</i>	<i>Vertical hydraulic conductivity (ft/d)</i>
Upper aquitard	1	0.005	0.005
Glasford aquifer	2	75	20
Low conductivity zone	2	1	1
Middle aquitard	3	0.0005	0.00025
Lower Glasford/upper Banner aquifer	4	120	20
Lower aquitard	5	0.0005	0.0004
Sankoty-Mahomet Sand aquifer	6	275	35
Low conductivity zone 1	6	75	20
Low conductivity zone 2	6	1	1
Lacustrine material and bedrock	7	inactive	inactive

Note: ft/d - feet per day

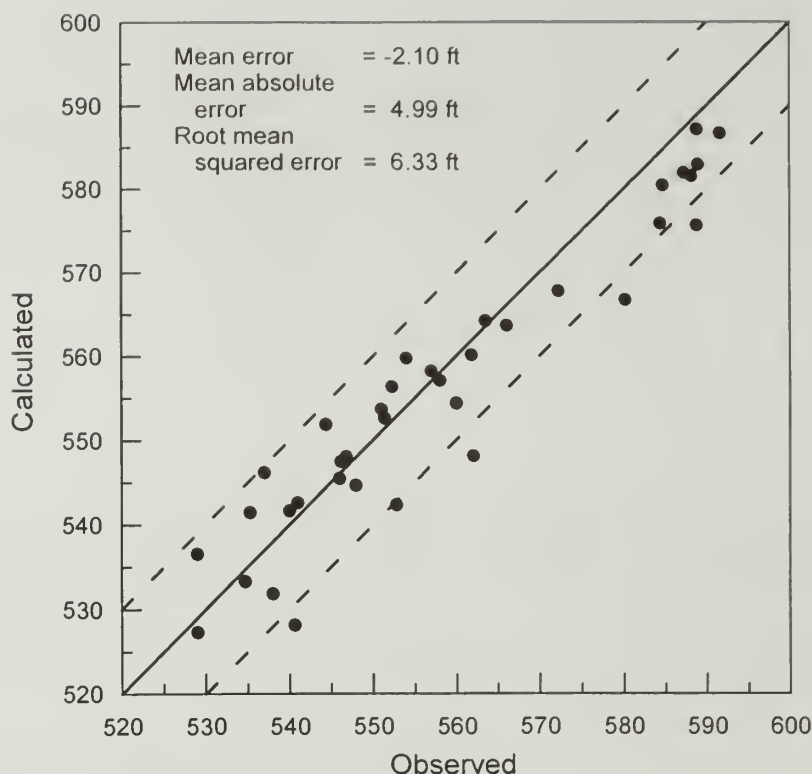


Figure 36. Calculated versus observed heads (in feet above mean sea level)

Valley to the system in the study area with direct leakage from the upper aquifers. To match the increased gradient in the model, ground-water levels were raised in the upper aquifer layers, which included simulating the loss of water from Sugar Creek to the Glasford aquifer layer. In addition, the hydraulic conductivity was lowered to 75 ft/d when there were rapid changes in the aquifer geometry along the bedrock valley walls and in the north-central part of T22N, R2W (figure 43), where there is only 12 feet of dirty sand and gravel. Lowering the hydraulic conductivity of the rest of the aquifer from the initial estimate of 335 ft/d to 275 ft/d also helped match the hydraulic gradients in the transition area; however, any further reduction caused gradients elsewhere in the model to be too steep. Because ground-water flow is largely horizontal, the model was not sensitive to changing the vertical hydraulic conductivity of the aquifer.

The underestimated heads (figure 38) continue westward to the Emden area south of the Hopedale bedrock high where flow begins to diverge to the southwest and to the northwest. Extending the Hopedale high a mile and a half to the west reduced this error by partially damming the flow, thus increasing water levels up gradient of the bedrock high. The error in predicted water levels also was reduced a small amount by increasing the recharge rate to 5 inches/year (figure 35) in the northern part of T21N, R4W where the Glasford aquifer sand is at the surface.

The model originally overestimated water levels around the Normal well field. The steep gradient in the potentiometric surface (figure 20) in T24N, R1E suggests that the mapped Sankoty-Mahomet Sand aquifer layer has a much lower hydraulic conductivity in the Danvers Bedrock Valley. In addition, the shallow gradient in T24N, R1W suggests there is very little water moving into the main body of the aquifer from this side valley. The model error was minimized by shutting off regional flow from the Danvers Bedrock Valley and by placing

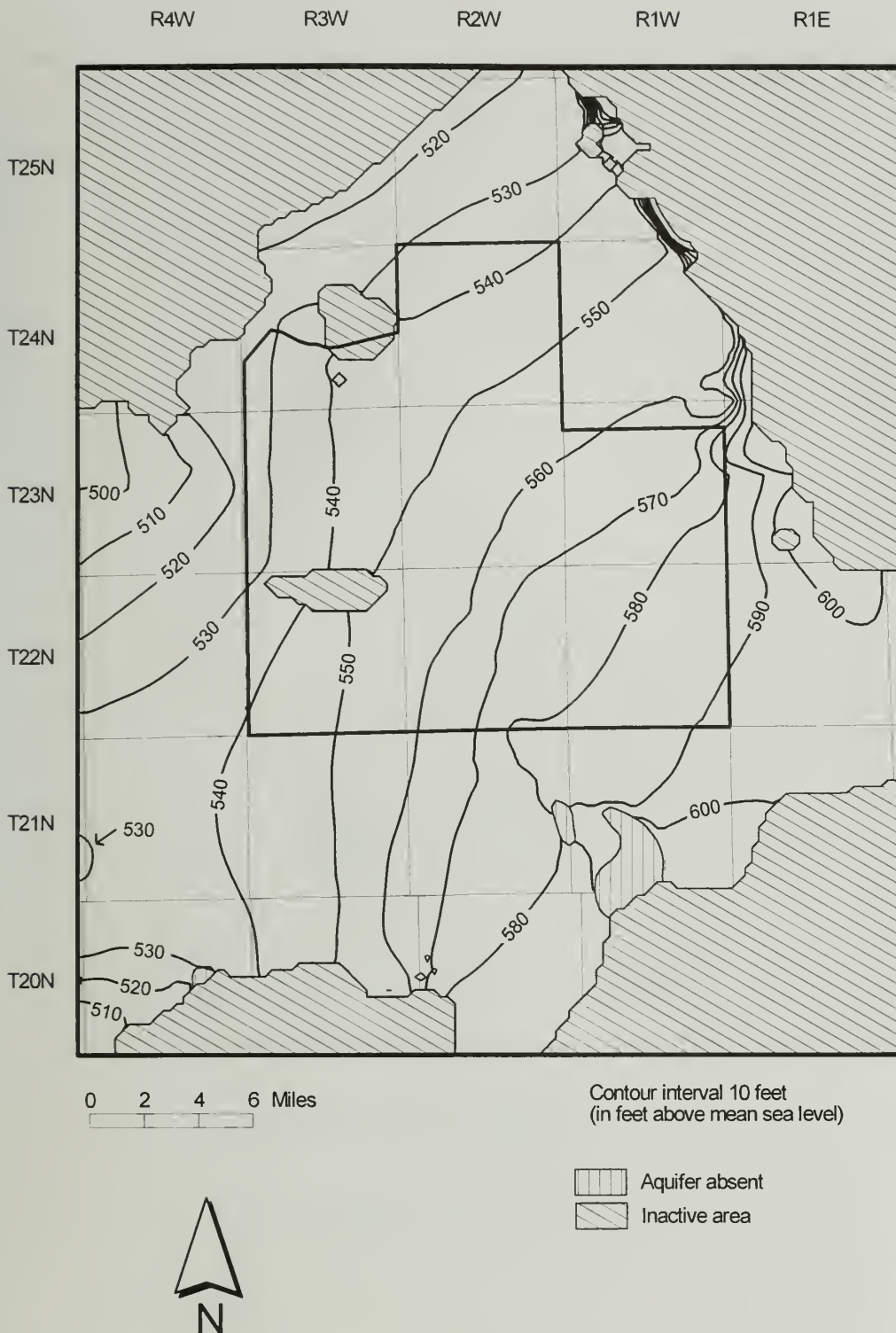


Figure 37. Modeled Sankoty-Mahomet Sand aquifer layer potentiometric surface
(numbers indicate contour elevation)

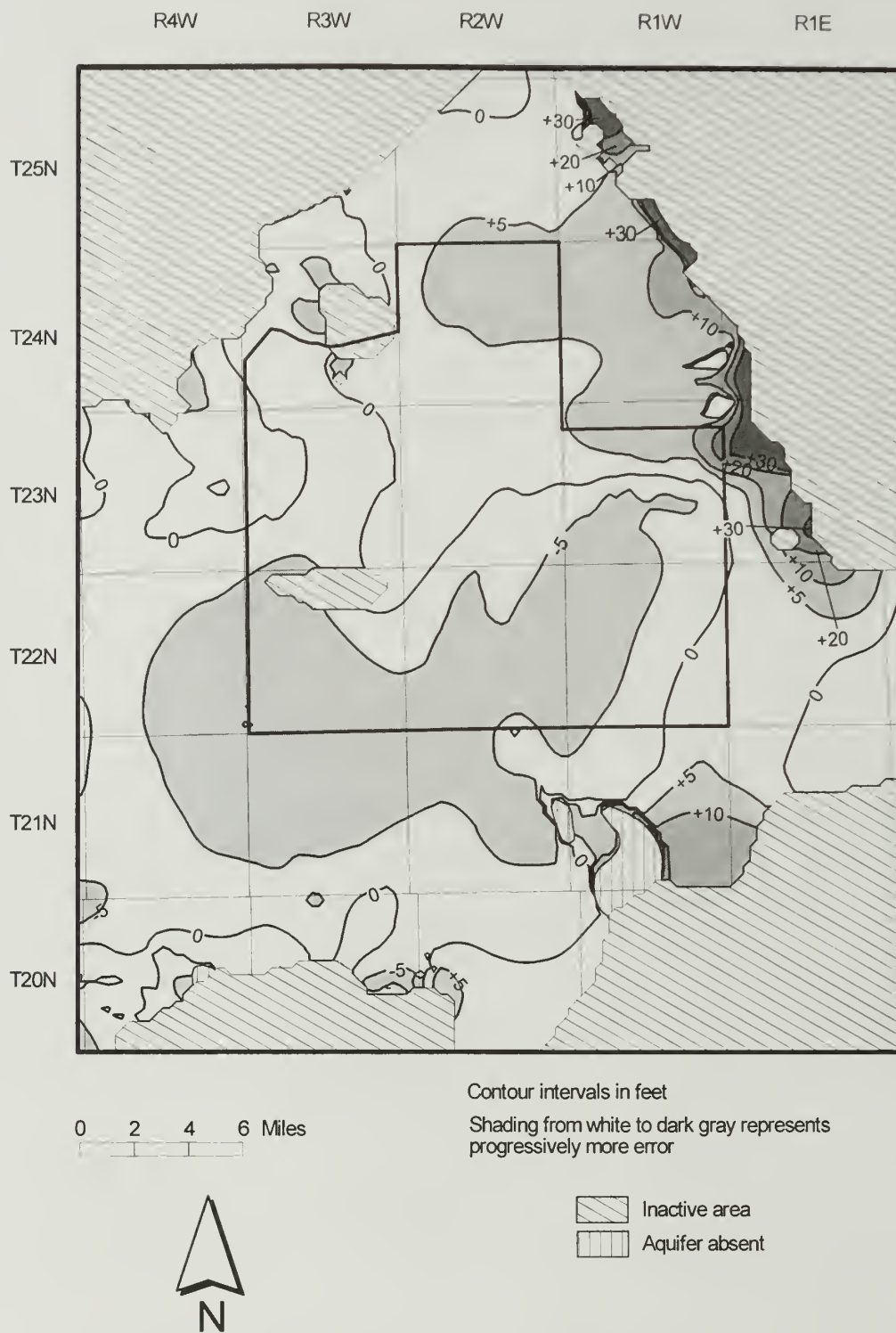


Figure 38. Error in modeled Sankoty-Mahomet Sand aquifer layer potentiometric surface (numbers indicate feet of error)

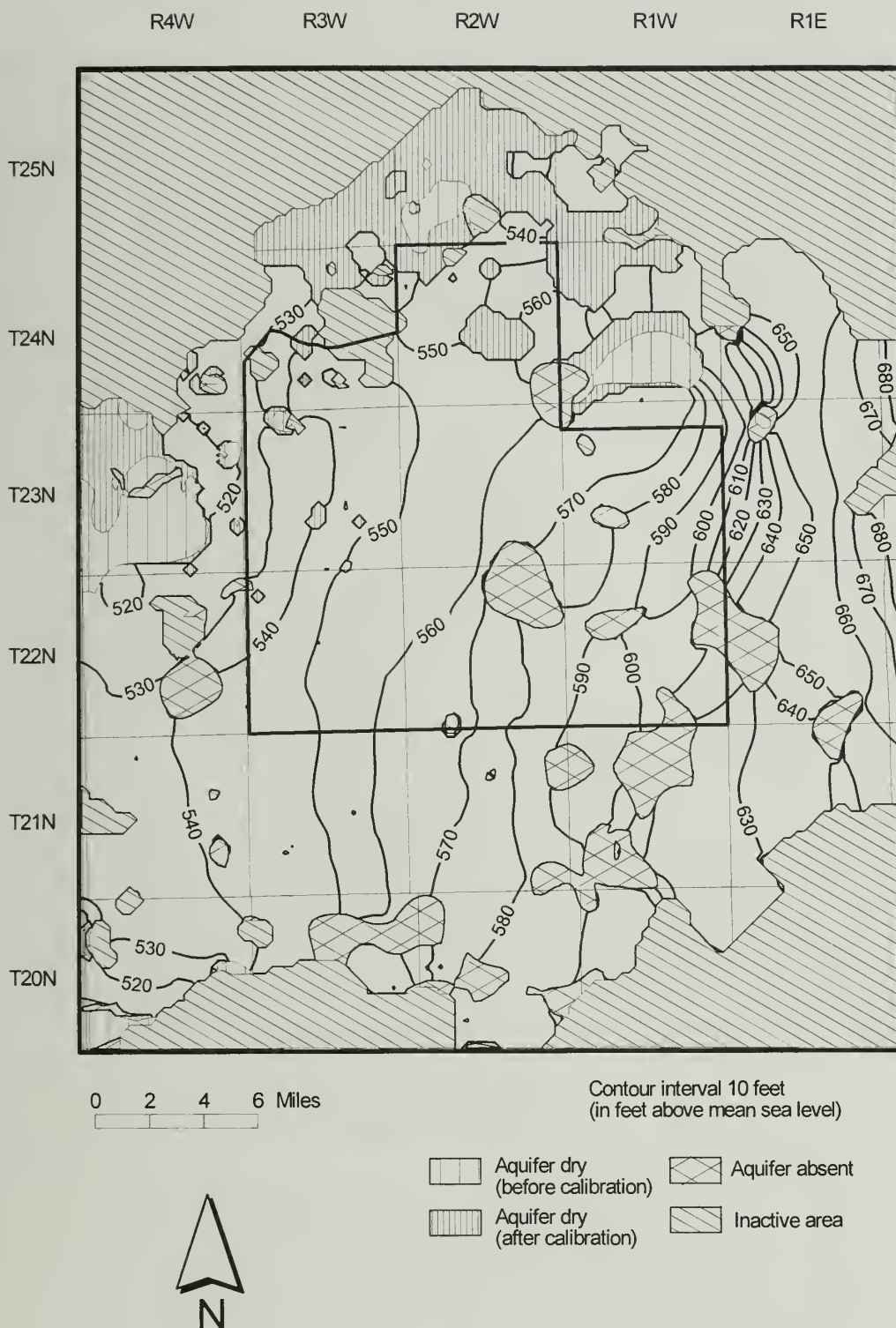


Figure 39. Modeled lower Glasford/upper Banner aquifer layer potentiometric surface (numbers indicate contour elevation)

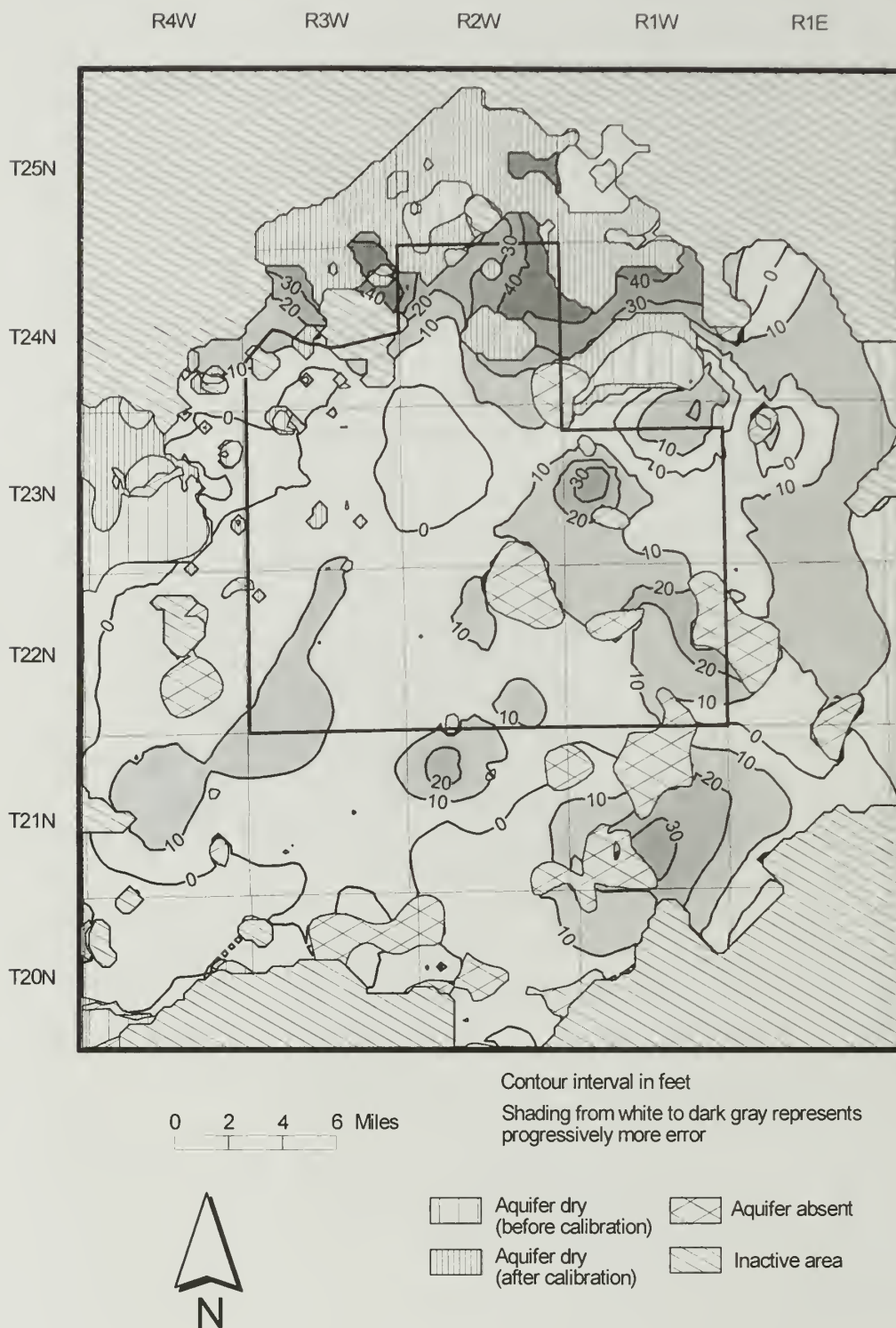


Figure 40. Error in modeled lower Glasford/upper Banner aquifer layer potentiometric surface (numbers indicate feet of error)

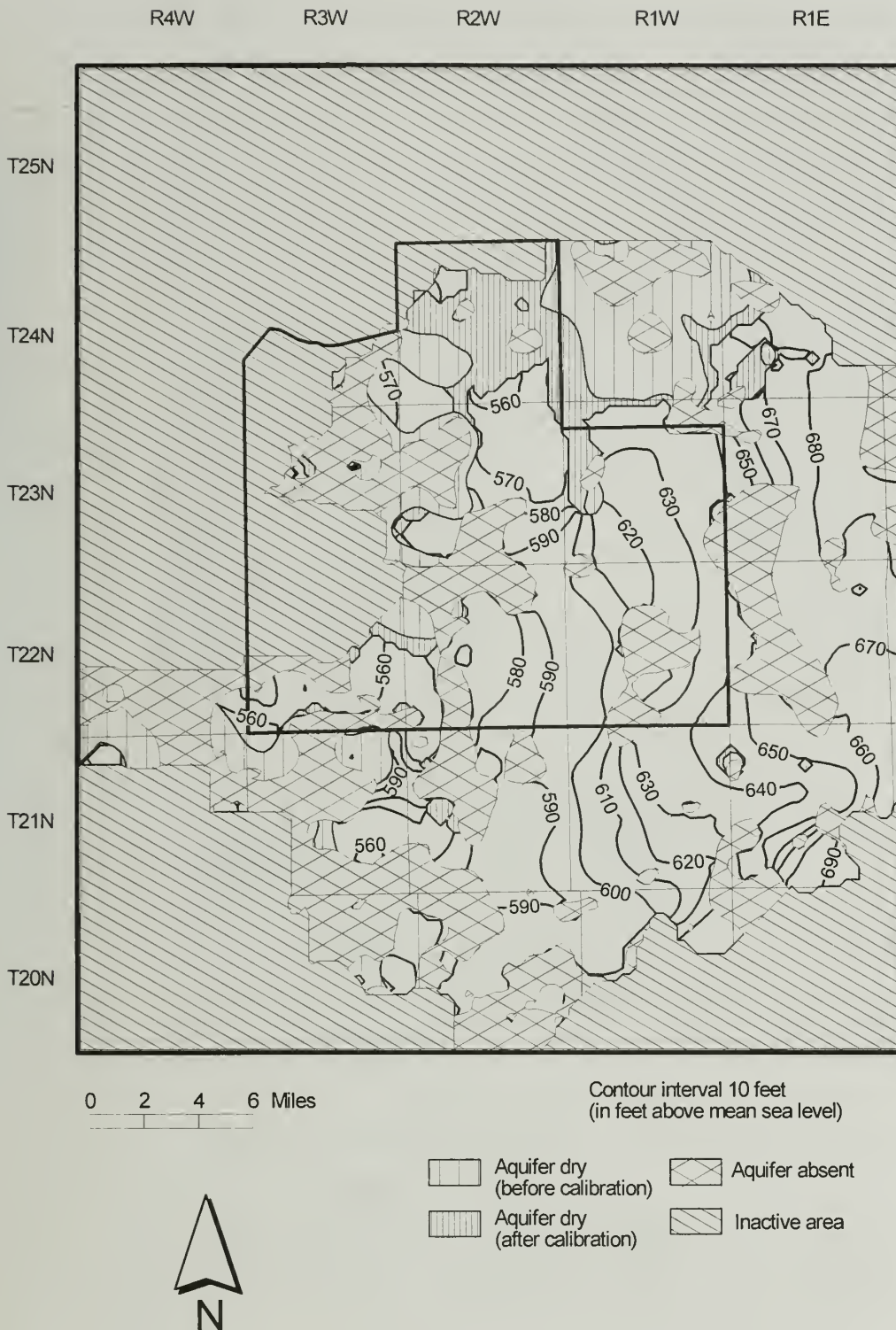


Figure 41. Modeled Glasford aquifer layer potentiometric surface
(numbers indicate contour elevation)

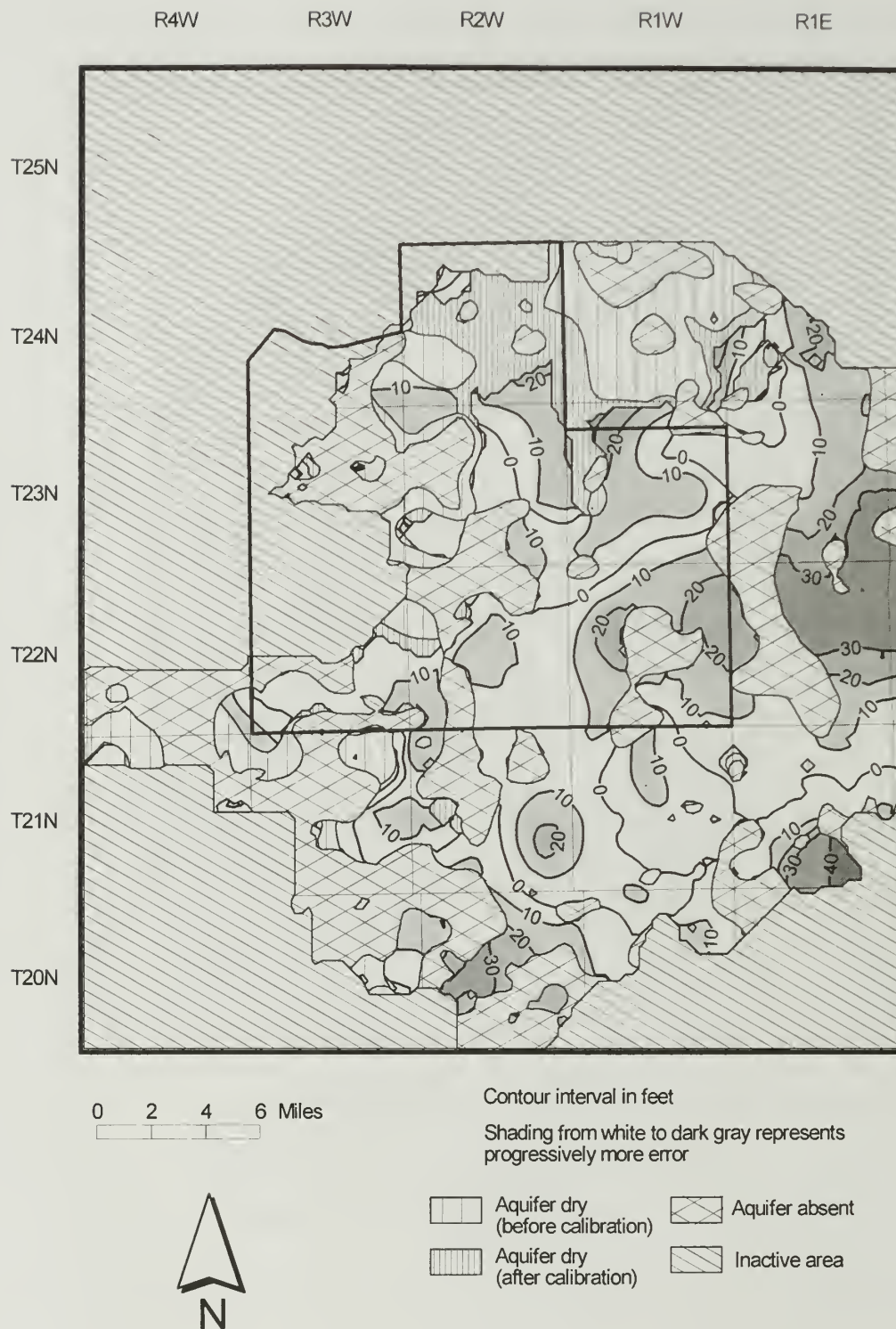


Figure 42. Error in modeled Glasford aquifer layer potentiometric surface (numbers indicate feet of error)

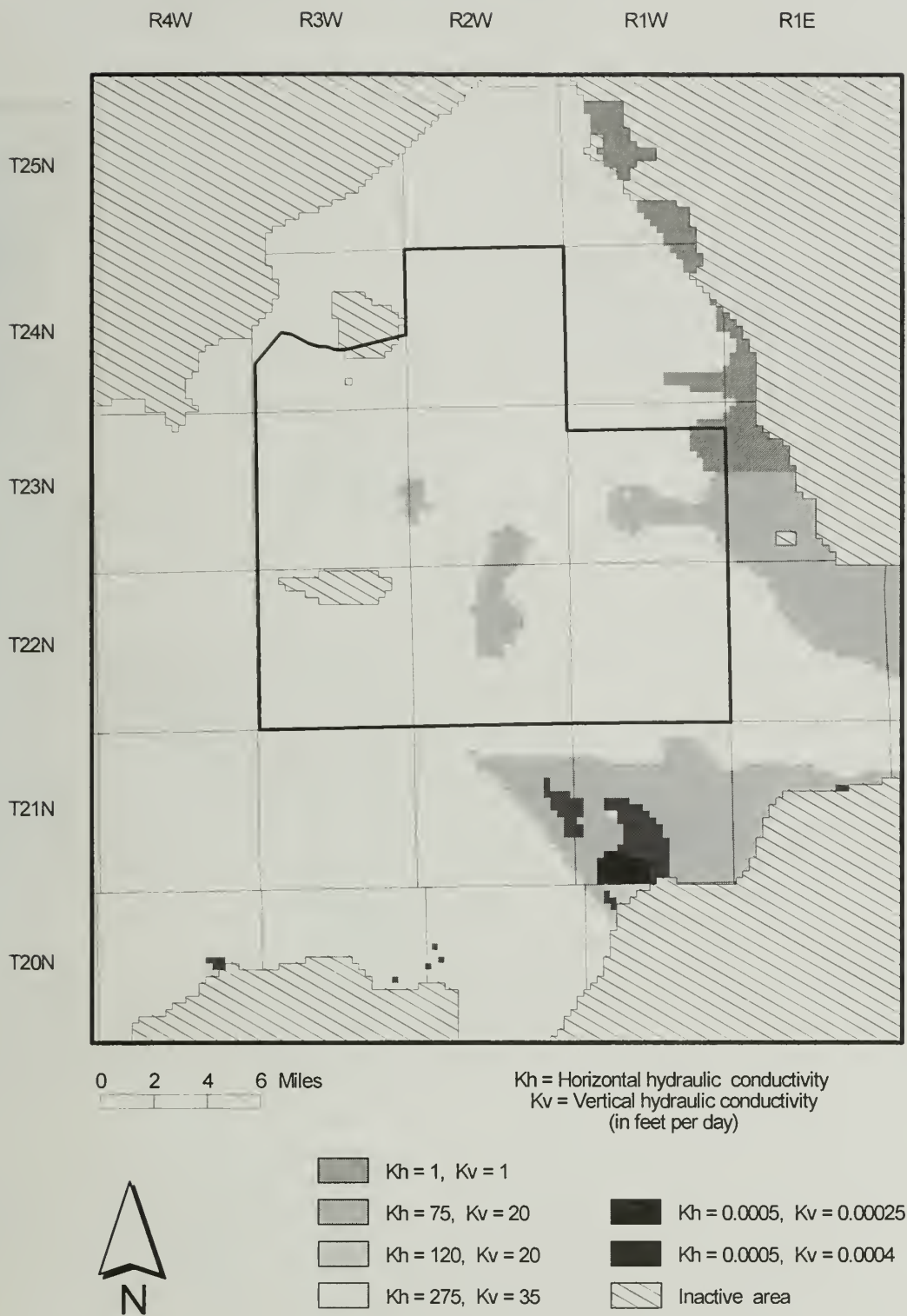


Figure 43. Hydraulic conductivity zones in model layer 6

some low conductivity zones around the well field to match the well field conceptualization put forth by Richards and Visocky (1982). More data from the township north of the Normal well field are needed to more accurately model this portion of the aquifer.

In the upper portion of the model, a significant amount of fine tuning was needed to keep the Glasford aquifer layer saturated and not have the water table in layer 1 go above land surface. The initial hydraulic conductivity of 170 ft/d produced ground-water levels that were too low and gradients that were too flat. This problem was solved by reducing the hydraulic conductivities of the lower Glasford/upper Banner and the Glasford aquifer layers to 120 ft/d and 75 ft/d, respectively. As in the lower aquifer, the model was not sensitive to changing the vertical hydraulic conductivity of the aquifer because flow is largely horizontal. The remaining areas with large errors in the upper two aquifers are in such areas as T22N, R1W (figures 40 and 42), where the aquifer layers are not continuous and there is insufficient geologic and water-level data to define the potentiometric surface accurately. In the northern portion of the model area, parts of the Glasford and lower Glasford/upper Banner model layers went dry. This is largely due to the drop off in the hydraulic head of the Sankoty-Mahomet Sand aquifer going toward the Illinois River and the direct connections between all three aquifer layers.

A problem occurred with the elevation of the water table exceeding the land surface elevation for which the upper aquitard could not transmit the recharge flux being added to the top of the model downward to the Glasford aquifer layer. To solve this problem, the hydraulic conductivity of the upper aquitard was increased by an order of magnitude to 0.005 ft/d. This higher value can be accounted for by the many thin sand deposits commonly reported on the drillers' logs that help transmit water through the aquitard or to localized discharge points along the smaller streams. When the Glasford aquifer layer was absent, the problem was solved by reducing recharge to either 0.20 inches/year or 0.35 inches/year and/or by assigning a hydraulic conductivity of 1.0 ft/d to layer 2 (figure 44).

As is always the case, the model could have benefitted from additional observation wells, especially outside the 260-square mile study area. A larger regional model could better define the hydraulic parameters if the aquifer changes from confined conditions at the eastern edge of the model to unconfined conditions at the western edge.

Ground-Water Flow Budget

The ground-water flow budgets at the different boundaries simulated in the model are shown on table 4. The values listed in each column are the net inflow or outflow for each of the listed components. The total gross values are higher because some components, such as the Mackinaw River, both gain and lose water. As expected with any steady-state model that has constant-head boundaries, the discrepancy between the volumes of water in and out of the model is very close to zero (0.03 percent).

The budget numbers show that 80 percent of the water coming into the model is from infiltration of precipitation at the land surface, 11 percent is from regional flow in the Mahomet Bedrock Valley to the east, and 8 percent is from river leakage. The budget also shows that 57 percent of the water leaves the model through discharge to the rivers, 33 percent goes to the regional flow to the north and to the west, and the remaining 10 percent goes to production wells. The amounts of discharge to the Mackinaw River (1,673,380 ft³/d) is within 8 percent of the estimated 1,800,000 ft³/d (13.5 mgd) gain calculated from the 7-day, 10-year low-flows (Singh and Adams, 1973). Likewise, loss of water along Sugar Creek above Hartsburg (286,371 ft³/d) is within the estimated range of 260,000 ft³/d to 350,000 ft³/d (1.9 mgd to 2.6 mgd) (Singh and Adams, 1973).

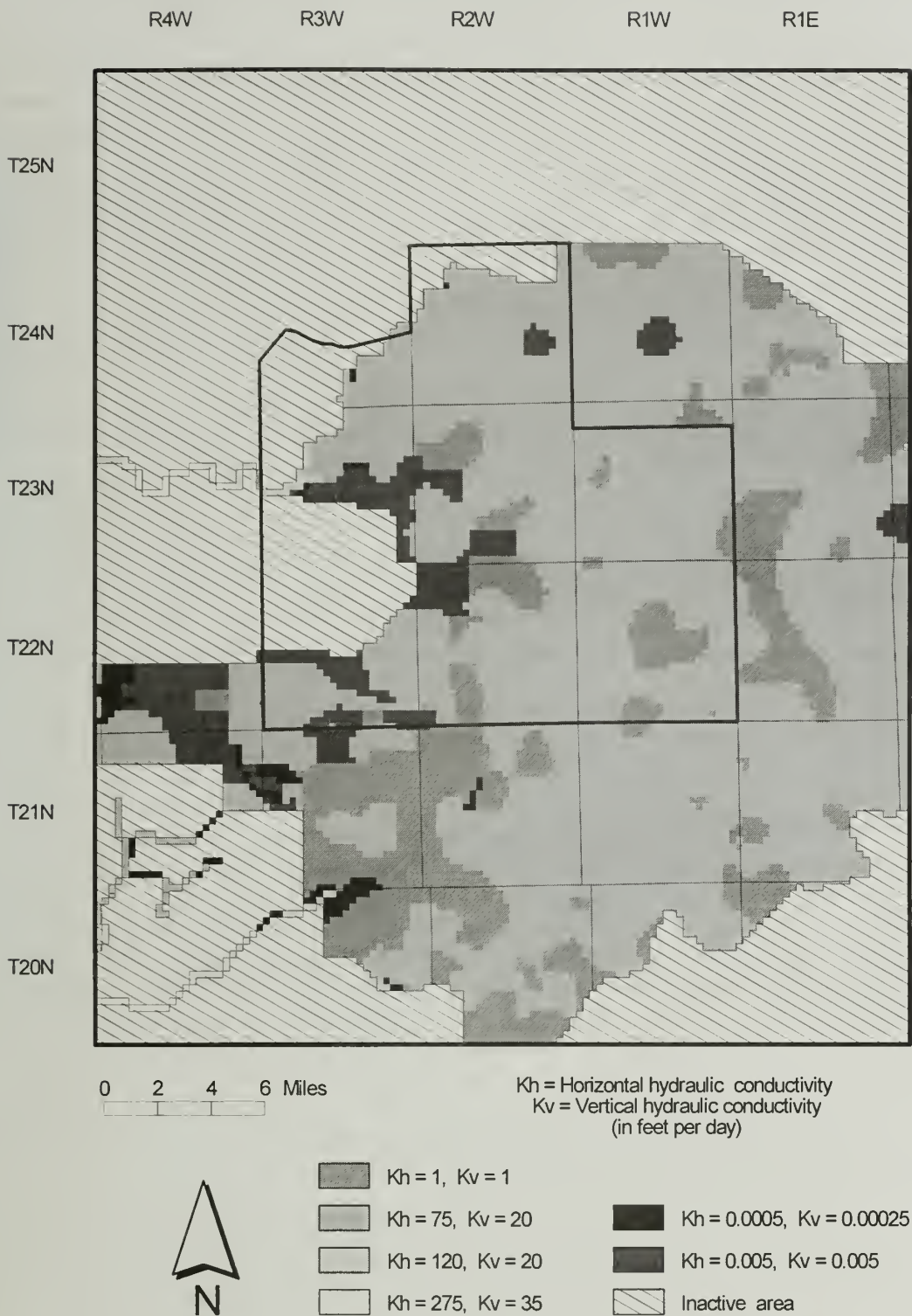


Figure 44. Hydraulic conductivity zones in model layer 2

Table 4. Net Ground-Water Flow Budgets of the Different Boundary Conditions Generated by the Calibrated Model

<i>Boundary condition</i>	<i>Inflow ft³/day (mgd)</i>	<i>Outflow ft³/day (mgd)</i>
Model edges	524,054 (3.92)	1,607,610 (12.0)
Mahomet and Kenny Bedrock Valleys	524,054 (3.92)	
Mackinaw Bedrock Valley		1,189,000 (8.89)
Western boundary		418,610 (3.13)
Streams	402,420 (3.01)	2,807,240 (21.0)
Mackinaw River		1,673,380 (12.5)
Little Mackinaw River	116,049 (0.87)	
Sugar Creek above Hartsburg	286,371 (2.14)	
Sugar Creek below Hartsburg		766,376 (5.73)
Kickapoo Creek		252,453 (1.89)
Prairie Creek		115,031 (0.86)
Production wells		495,026 (3.70)
Infiltration from precipitation	3,984,916 (29.8)	
Net budget	4,911,390 (36.7)	4,909,876 (36.7)

Notes: ft³/day - cubic feet per day
mgd - million gallons per day

Ground-water flow budgets can be calculated for portions of the Sankoty-Mahomet Sand aquifer layer, but the total budget for the entire aquifer layer is difficult to calculate because of complicated flow paths. For the southeastern third of the Sankoty-Mahomet Sand aquifer layer (townships 20N to 23N and the southern half of 24N and ranges 2W to 1E), the amount of water entering the Sankoty-Mahomet Sand aquifer layer from the layer above was computed to be 2,126,500 ft³/d (15.9 mgd). Fifty percent of this water came from the lower Glasford/upper Banner aquifer layer through areas in which the lower aquitard layer is absent. These areas comprise only 4 percent of this area. The other 50 percent is leakage through the lower aquitard, which comprises the other 96 percent of this area. These numbers underscore the importance of the direct connections between the Sankoty-Mahomet Sand aquifer layer and the overlying aquifer layers.

Ground water entering the model from the Mahomet and Kenny Bedrock Valleys accounts for 11 percent of the net total inflow. Assuming a chloride concentration of 71 milligrams per liter (mg/l) for this inflow and 3.5 mg/l for ground water in the shallower aquifers (Herzog et al., 1995a), a dilution trend can be computed from the budget in the model. Initially, the dilution will progress slowly to around 30-35 mg/l as ground water moving through the aquifer is mixed with leakage only through the lower aquitard. However, the dilution should greatly increase as flow in the aquifer moves past the areas of connection to the shallow aquifers, and the chloride concentration should start to drop below 20 mg/l. The chloride value for the Sankoty-Mahomet Sand aquifer layer for the entire model area should average about 10 mg/l, assuming that all of the infiltration and stream loss

(89 percent of the total inflow) entering the shallow aquifers has a chloride concentration of 3.5 mg/l. This dilution trend computed from the model budget matches closely with the chloride concentration map (figure 23) and suggests that the model corroborates field data. In the future, a solute transport simulation could be added to the model to simulate the mixing pattern directly.

Model Uniqueness

A ground-water flow model generally is not considered to provide a unique solution, meaning that more than one combination of model parameters could produce the same hydraulic head distribution. For example, if the transmissivity of a modeled aquifer is doubled, the same hydraulic head and gradients could be maintained if the recharge rate is substantially increased. However, the resulting model may have twice as much water moving through it, which may produce unrealistically high ground-water discharge to streams. Fortunately for this study, there are water-level data for the streams and the aquifers, flow budget information from the streams, and chloride concentration data for the Sankoty-Mahomet Sand aquifer to calibrate the model against. The fact that this model closely simulates these three independent sets of field data provides assurance that the model can make reasonably reliable and realistic scenarios.

The elevation of the major streams and the water levels along the model boundaries constrains the basic configuration of water levels closely enough to have a mean absolute error of about 20 feet with any reasonable aquifer hydraulic conductivity and no infiltration from precipitation. The proper combination of hydraulic conductivities and the amount of infiltration from precipitation reduces most of this error to an acceptable level and produce a flow budget that approximates the observed data. The most important hydraulic conductivity is that of the Sankoty-Mahomet Sand aquifer layer, which is well known for portions of the aquifer at which aquifer tests have been made.

An uncertain quantity is the amount of leakage from the overlying layers to the Sankoty-Mahomet Sand aquifer layer. Leakage controls the height of the water level rise in the middle of the model. Values for leakage are not put into the model. Instead, the model calculates leakage based on the head differences between aquifer layers, how well the different aquifer layers are connected, the hydraulic conductivities of the upper two aquifer layers and the aquitards, and, most important, the amount of infiltration from precipitation. Because the two upper aquifer layers do not have significant discharge areas that are independent of the Sankoty-Mahomet Sand aquifer, almost all of the infiltration from precipitation coming into the model eventually reaches the Sankoty-Mahomet Sand aquifer. Therefore, changing the infiltration rate has a direct effect on the amount of water-level error in the Sankoty-Mahomet Sand aquifer layer. For example, reducing the infiltration rate by 60 percent to 0.00007 ft/d (0.3 inches/year) caused the mean absolute error in the predicted ground-water levels at the observation wells in the study area to jump from about 5 to 14 feet.

The effect of the uncertainty of the hydraulic conductivities of the upper aquifer and aquitard layers was reduced because there are two additional water level maps to calibrate against, and the proper amount of infiltration has to be passed through as leakage to the Sankoty-Mahomet Sand aquifer layer. The most sensitive parameter in the upper layers is the location of gaps in the middle and lower aquitards. Most of the water is transferred into the Sankoty-Mahomet Sand aquifer layer through these areas of hydraulic connection.

Regional Well-Field Scenarios

Four sites were selected for testing a simulated well field, although many other sites could have been selected. The sites (figure 45) were principally selected by examining the thickness of the Sankoty-Mahomet Sand aquifer layer. The most likely locations for a well field that would minimize impacts to the aquifer and existing wells are those in which the aquifer is the thickest and/or connected to an overlying aquifer. Proximity to

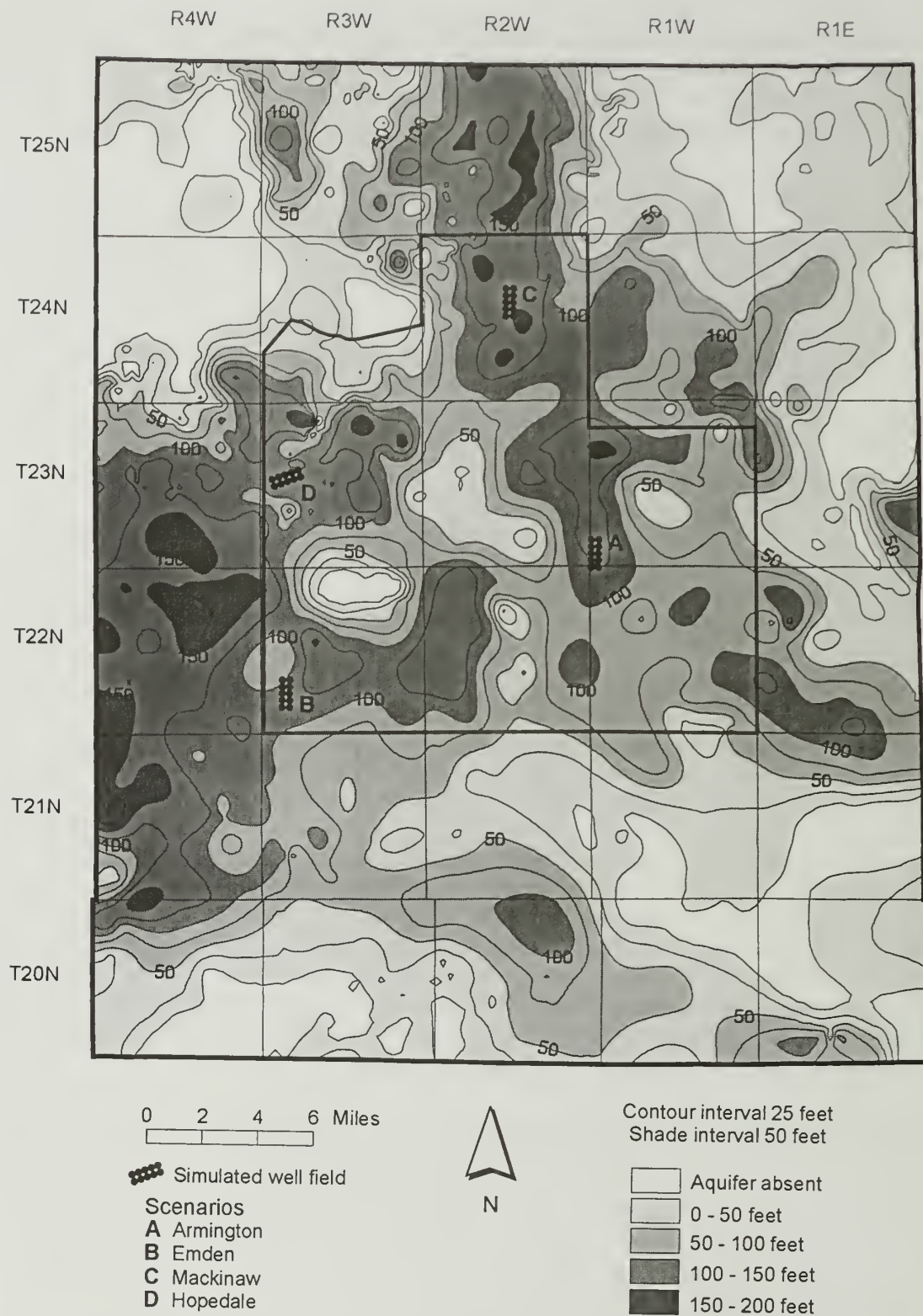


Figure 45. Locations of the four pumping scenarios tested with the model

the two aquifer tests conducted for this study with the most reliable hydraulic property data was also a factor in selecting two of the sites. All locations were placed within the 260-square mile study area. Each was given the name of a town near the site for easy reference on a map: Armington, Emden, Mackinaw, and Hopedale.

Each scenario modeled a well field, or group of pumping wells, that produced 15 mgd continuously. All of the simulated wells were screened only in the Sankoty-Mahomet Sand aquifer layer. Though the situations modeled are hypothetical, the setup and conditions put into the model were made as realistic as possible. Ten wells, each pumping 1,040 gallons per minute (gpm), were required to pump the 15 mgd in the simulated well fields. The wells were spaced one-quarter mile apart in a 5 x 2 well array. This design was based on realistic pump sizes and the drawdowns experienced during the aquifer tests. An engineering consultant (Farnsworth and Wylie, personal communication, 1998) agreed that this well-field design was realistic for the study area. The model was set to run to steady-state conditions, meaning that, if conditions in the aquifer system remained unchanged (recharge, pumpage, etc.), the model results would approximate a flow system in equilibrium. In other words, these results would be the maximum changes in water levels one would expect to see.

The outputs of the model are potentiometric surface elevations at each grid node for each layer under pumping conditions. For the three aquifer layers, these elevations (water levels) were subtracted from the potentiometric surface maps for the nonpumping conditions to develop maps of the change in water level due to pumpage, which are referred to as drawdown maps.

Armington Scenario

The location of the wells in this scenario are in Sections 30 and 31, T23N, R1W. The drawdown in the potentiometric surface of the Sankoty-Mahomet Sand aquifer layer from simulated pumpage at the Armington site is shown in figure 46. At this location, the Sankoty-Mahomet Sand aquifer layer is a thick, narrow, north-south band, and the layer is fairly thin both east and west of the simulated well field. These areas of changing aquifer thickness reduce the amount of ground water available to the well field by affecting the transmissivity of the aquifer and because of the reduced areal extent. The drawdowns at this site are greater than those of the other scenarios and extend over a larger area. In the simulation, the drawdown at the well field was approximately 55 feet. This value is an average over the 40-acre grid cell, and the model does not predict the drawdown in water levels in the pumping wells themselves. The extent of the 10-foot drawdown contour was as much as 15 miles from the well field and extended to the edge of the aquifer layer over a large portion of the eastern side of the model area. The drawdown also extended to the eastern edge of the model boundary along the Mahomet Bedrock Valley, causing the model to artificially supply more water than it should because of constant-head conditions. To enable a better simulation of ground-water levels in the study area, the constant heads were lowered by 10 feet in the Sankoty-Mahomet Sand aquifer layer and by 3 feet in the lower Glasford/upper Banner aquifer layer.

The potentiometric surface of the lower Glasford/upper Banner aquifer had more than 40 feet of drawdown as a result of pumpage from the Sankoty-Mahomet Sand aquifer at this site (figure 47). This is essentially the same amount of drawdown as was seen in the Sankoty-Mahomet Sand aquifer, a result of hydraulic connections between the aquifers and the amount of ground water available from the lower Glasford/upper Banner aquifer. The connections between the two aquifer layers allowed ground water in the lower Glasford/upper Banner aquifer to recharge the Sankoty-Mahomet Sand aquifer. Pumpage in the Sankoty-Mahomet Sand aquifer reduced the potentiometric surface at the well field, which increased the hydraulic gradient between the two aquifers and increased the volume of ground water recharged through the connection between the aquifers.

The Glasford aquifer model layer also was influenced by pumpage from the Sankoty-Mahomet Sand aquifer layer at this location (figure 48), with drawdown in the potentiometric surface of about 20 feet. The areas

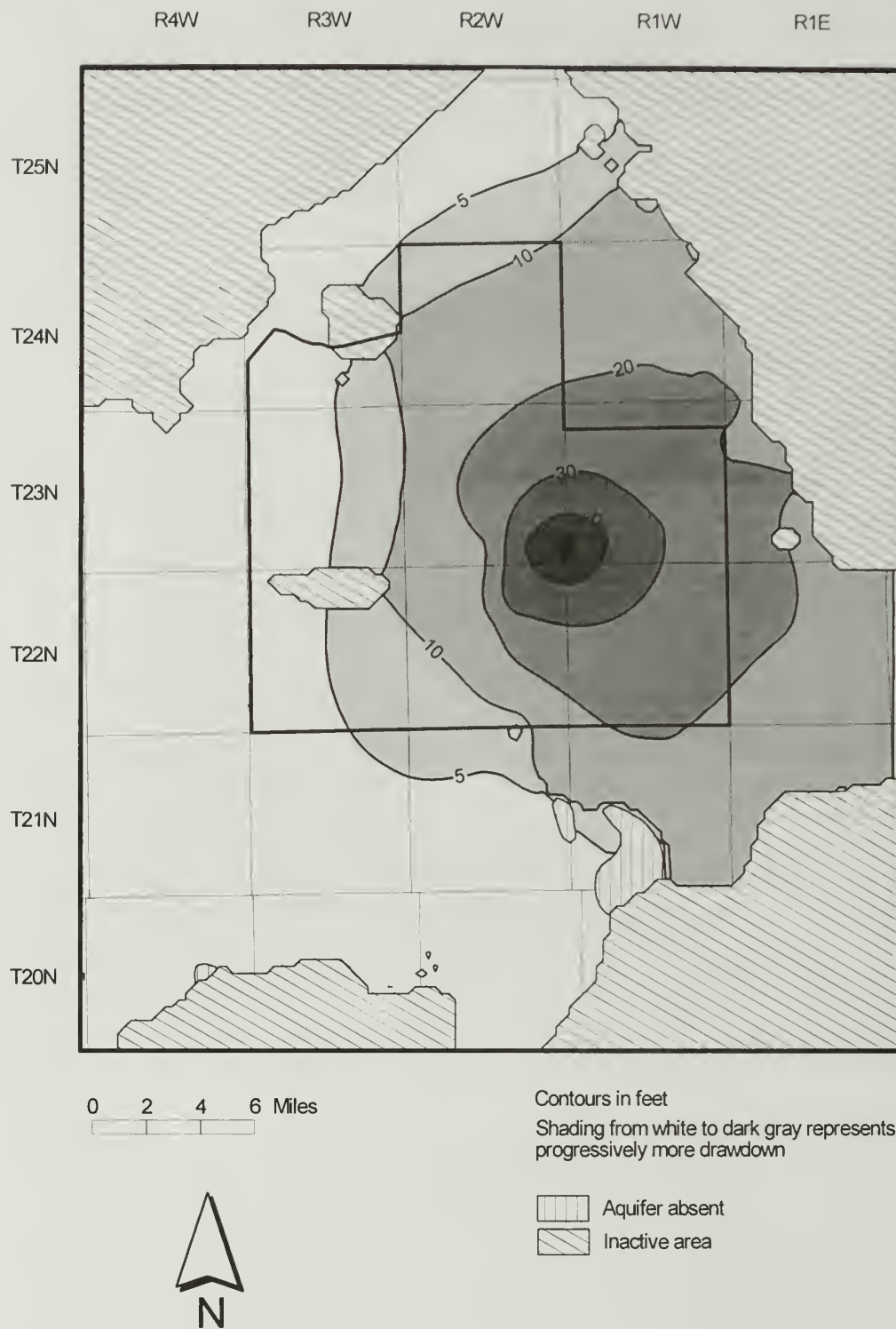


Figure 46. Drawdown in the Sankoty-Mahomet Sand aquifer layer in the Armington scenario (numbers indicate feet of drawdown)

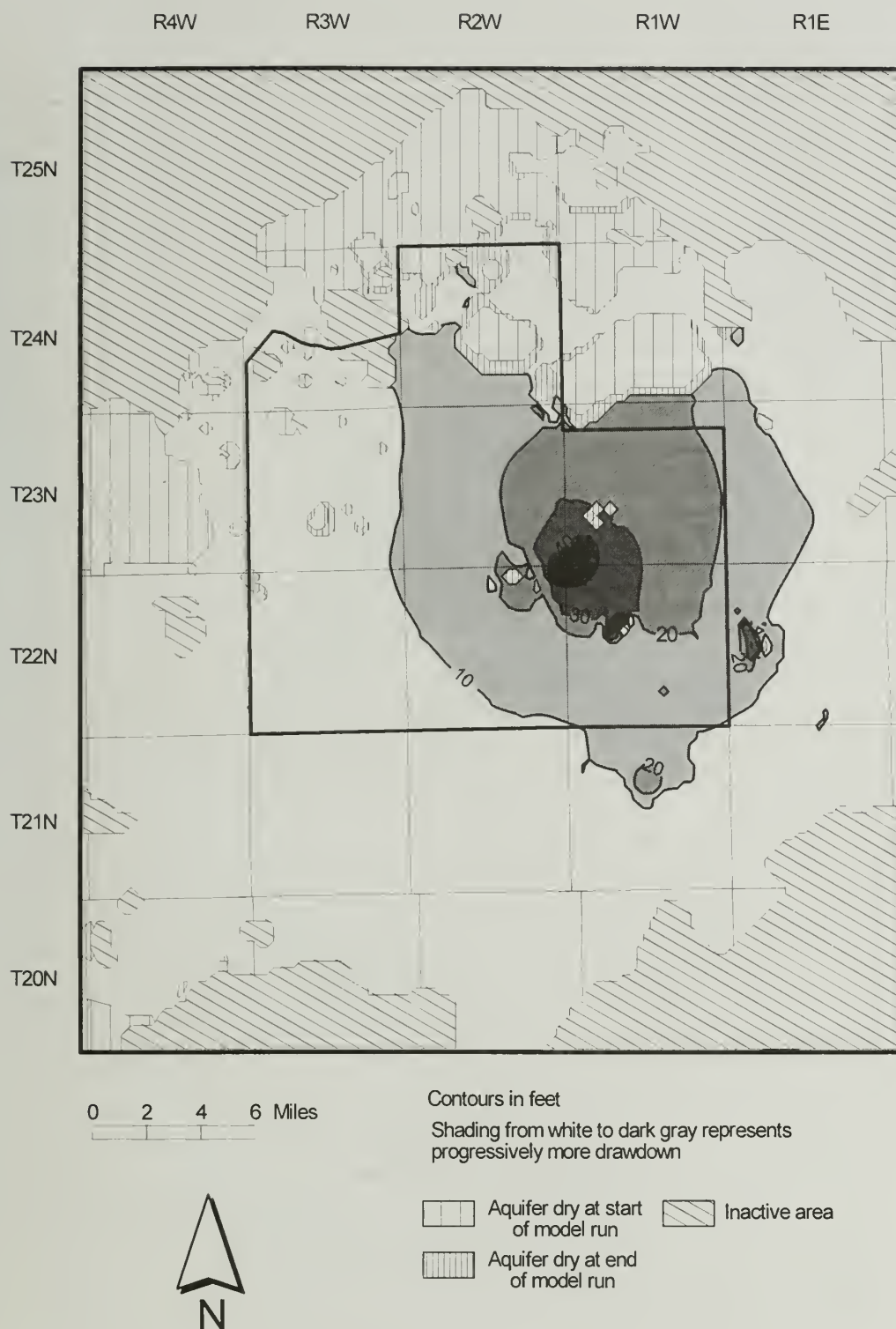


Figure 47. Drawdown in the lower Glasford/upper Banner aquifer layer in the Armington scenario (numbers indicate feet of drawdown)

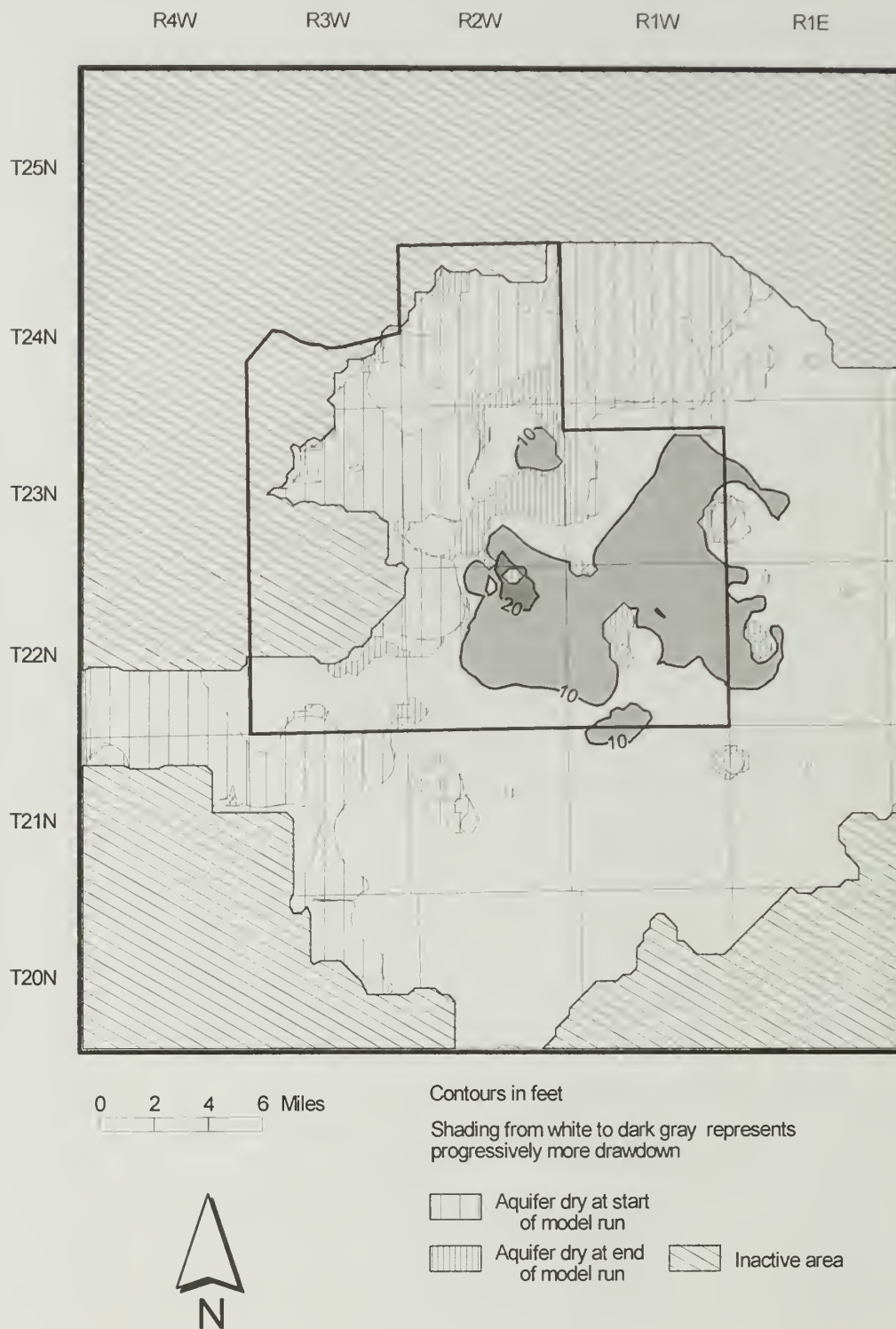


Figure 48. Drawdown in the Glasford aquifer layer in the Armington scenario (numbers indicate feet of drawdown)

of greatest drawdown are where there is a large area of connection between the Glasford aquifer layer and the lower Glasford/upper Banner aquifer layer. This connection, as with the connection between the lower two aquifer layers, enables the Glasford aquifer to readily recharge the lower Glasford/upper Banner aquifer as its water level declines. The area of the 10-foot drawdown contour extends to the northeast 8-10 miles. This may be indicating the mutual influence/interference from the Normal well field. As the hydrologic evaluation showed, the Glasford aquifer was unsaturated in nearby parts of this area. The model also showed that these unsaturated aquifer areas may be expanded by a well field in the Armington area.

In the model, the ground-water flow system achieves a new equilibrium whereby the pumpage at the well-field site is offset by a combination of increased inflow and reduced outflow for the other hydrologic components of the system. Table 5 provides the ground-water flow budget of the model for the Armington scenario. Because the system has reached steady-state, the overall inflow and outflow are nearly equal. Therefore, the differences between the individual components in tables 4 and 5 describe the effect of the pumpage at the well field. For instance, the water leaving the model through the model edges was reduced from 1,607,610 ft³/day (12.0 mgd) (table 4) to 1,015,650 ft³/day (7.6 mgd) (table 5).

Table 5. Net Ground-Water Flow Budgets of the Different Boundary Conditions Generated for the Armington Scenario

<i>Boundary condition</i>	<i>Inflow ft³/day (mgd)</i>	<i>Outflow ft³/day (mgd)</i>
Model edges	916,723 (6.86)	1,015,650 (7.60)
Mahomet and Kenny Bedrock Valleys	916,723 (6.86)	
Mackinaw Bedrock Valley		737,440 (5.52)
Western boundary		278,210 (2.08)
Streams	945,224 (7.07)	2,330,541 (17.4)
Mackinaw River		1,280,266 (9.58)
Little Mackinaw River	214,256 (1.60)	
Sugar Creek above Hartsburg	730,965 (5.47)	
Sugar Creek below Hartsburg		708,731 (5.30)
Kickapoo Creek		231,986 (1.74)
Prairie Creek		109,558 (0.82)
Production wells		2,500,336 (18.7)
Infiltration from precipitation	3,985,811 (29.8)	
Net budget	5,847,758 (43.7)	5,846,527 (43.7)

Notes: ft³/day - cubic feet per day
mgd - million gallons per day

Emden Scenario

The location of the wells in the Emden scenario are in Sections 20 and 29, T22N, R3W. The drawdown in the potentiometric surface of the Sankoty-Mahomet Sand aquifer layer from simulated pumpage at the Emden site is shown in figure 49. At this location the Sankoty-Mahomet Sand aquifer layer is fairly thick (generally >100 feet) in all directions except for the bedrock high to the northeast near Hopedale, which pinches out the aquifer. The aquifer test conducted near Emden indicated the bedrock high would influence drawdown in the area, though not significantly. In figure 49, the drawdown extends toward the bedrock high, as shown by the 20-foot drawdown contour. The drawdown at the well field was approximately 35 feet, although this value will be greater in the pumping wells themselves. The extent of the 10-foot drawdown contour was only about 6 miles from the well field. Drawdown reached the western edge of the model, which was accommodated by removing some of the constant-head cells and by lowering the head 3 feet at others.

More than 25 feet of drawdown occurred in the potentiometric surface of the lower Glasford/upper Banner aquifer layer in this scenario. The largest drawdowns were at locations where the aquifer was directly connected to the Sankoty-Mahomet Sand aquifer layer (figure 50). In this case, part of this aquifer layer became unsaturated north of the site. The drawdown was fairly uniform in all directions, which was expected because of the continuity of the lower Glasford/upper Banner aquifer layer in this area.

The Glasford aquifer layer had drawdowns in the potentiometric surface of more than 30 feet as a result of pumpage from the Sankoty-Mahomet Sand aquifer layer. In addition, a significant portion of the Glasford aquifer layer became unsaturated as a result of this pumpage (figure 51). Figure 18 shows a large area of physical connection between the upper two aquifers. Some of the ground-water levels for the Glasford aquifer layer in this area are based on stream data. In an aquifer discharging to a stream, the water levels would be lowest at the stream. In general, because the water levels are actually higher in areas adjacent to the stream, the unsaturated areas are probably more extensive in the model simulation than would occur in the field.

The connections between aquifer layers at this site are complex. At the well field and north of the well field, the Glasford aquifer layer and the lower Glasford/upper Banner aquifer layer are connected. A quarter mile south of the well field, the connection between the upper aquifer layers disappears, and a connection between the lower Glasford/upper Banner aquifer layer and the Sankoty-Mahomet Sand aquifer layer appears.

Mackinaw Scenario

The location of the wells in this scenario are in Sections 10 and 15, T24N, R2W. Drawdown in the potentiometric surface of the Sankoty-Mahomet Sand aquifer layer from simulated pumpage at the Mackinaw site is shown in figure 52. In the simulation, the drawdown at the well field was approximately 45 feet. The extent of the 10-foot drawdown contour was as much as 12 miles from the well field; it also extended to the aquifer boundary over much of the area east of the well field. The drawdown north of the well field extended to the north edge of the model in the Mackinaw Bedrock Valley. Because 8.89 mgd of ground water is allowed to flow out of the model along this edge, adjusting the constant-head values to match the drawdown was not straightforward. Removing the constant-head conditions altogether produced almost the same ground-water levels as leaving them set to 517, meaning that flow to the north is largely diverted by this pumping scenario. Better prediction of the 5-foot and 10-foot drawdown contours in the potentiometric surface north of Mackinaw would require extending the model beyond the Morton well field to the Illinois River.

The drawdown in the potentiometric surface of the lower Glasford/upper Banner aquifer layer is shown in figure 53. In this scenario, the aquifer exhibits the same properties as the other scenarios; areas of greater drawdown correspond to locations at which the aquifer is hydraulically connected to the Sankoty-Mahomet Sand

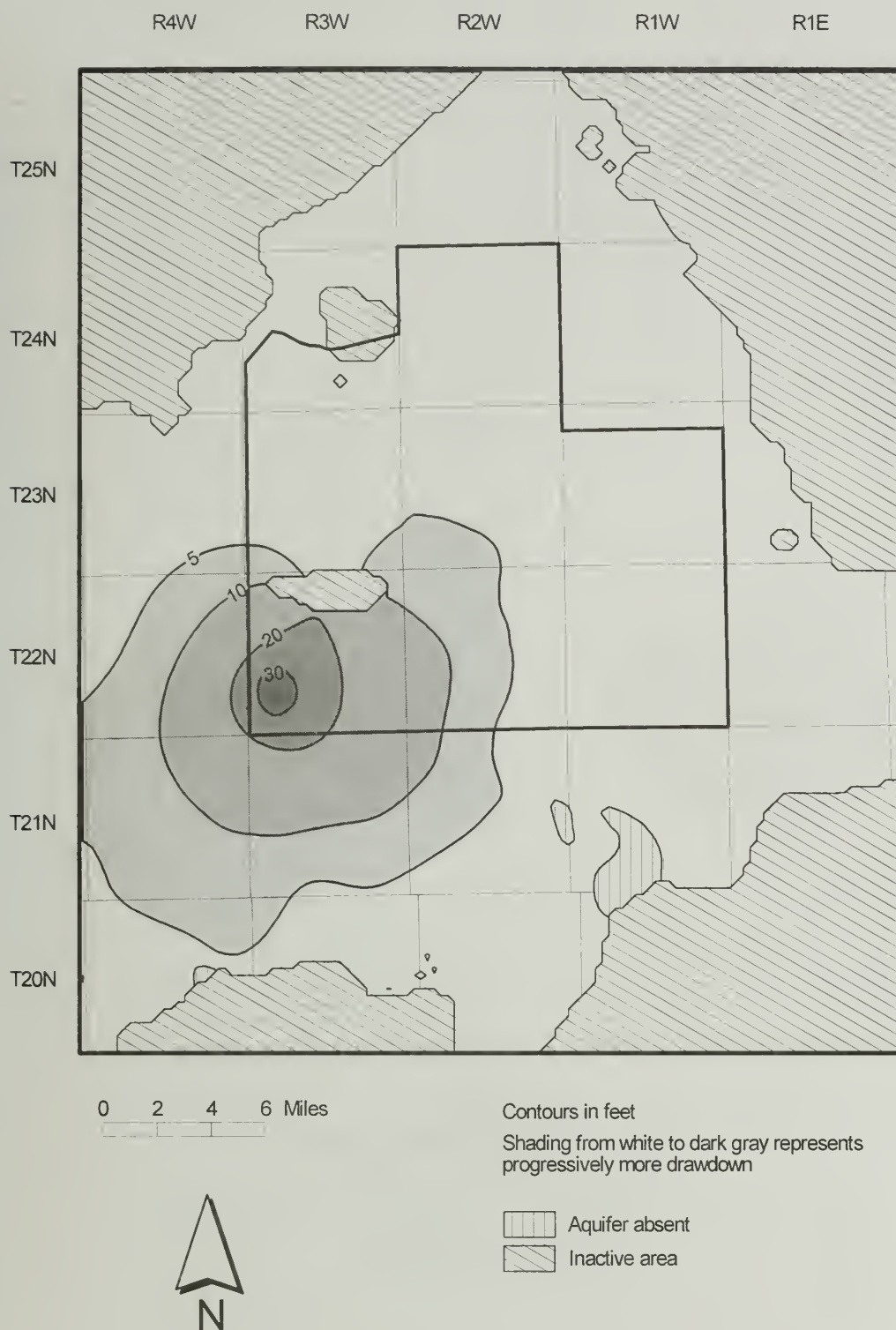


Figure 49. Drawdown in the Sankoty-Mahomet Sand aquifer layer in the Emden scenario (numbers indicate feet of drawdown)

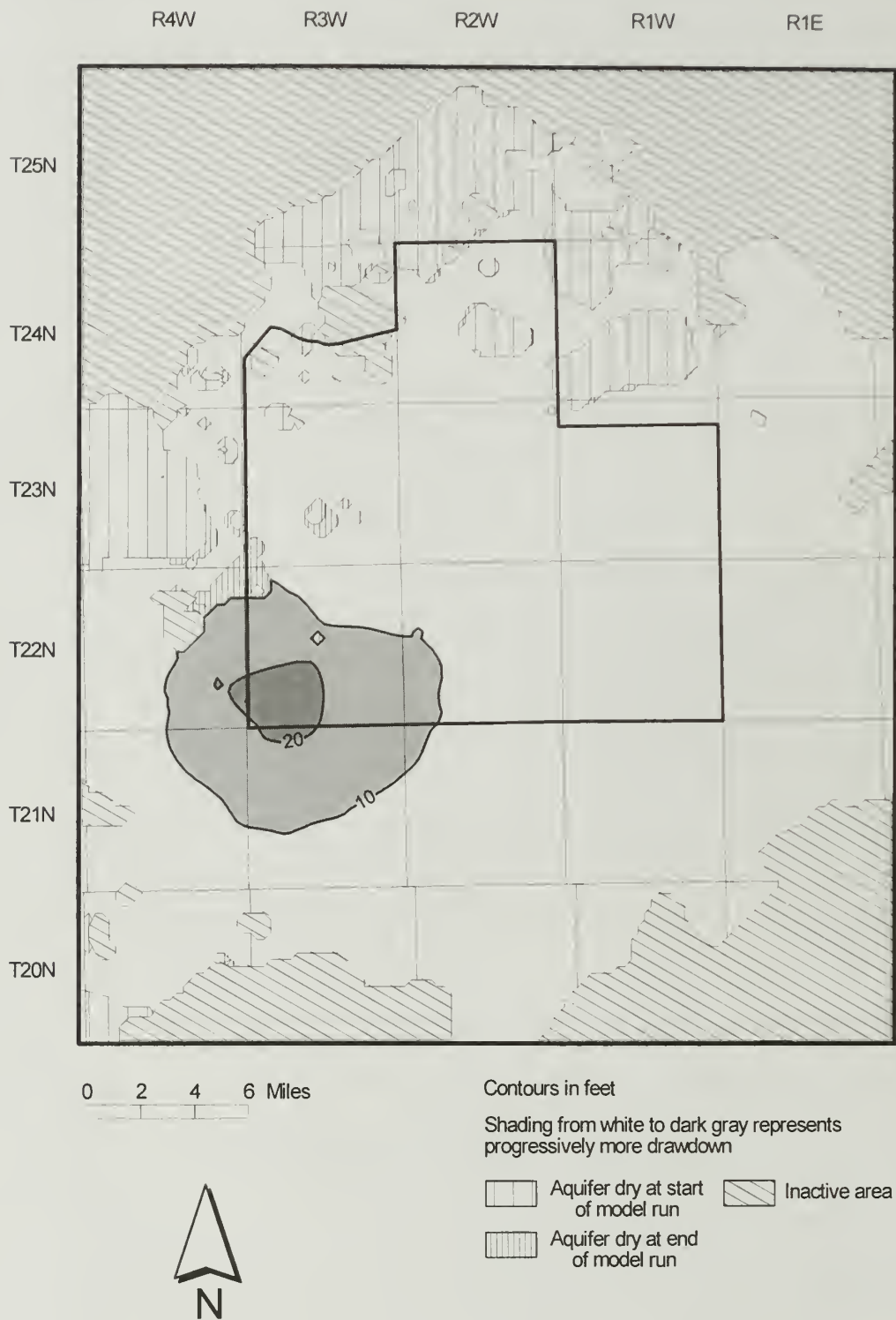


Figure 50. Drawdown in the lower Glasford/upper Banner aquifer layer in the Emden scenario (numbers indicate feet of drawdown)

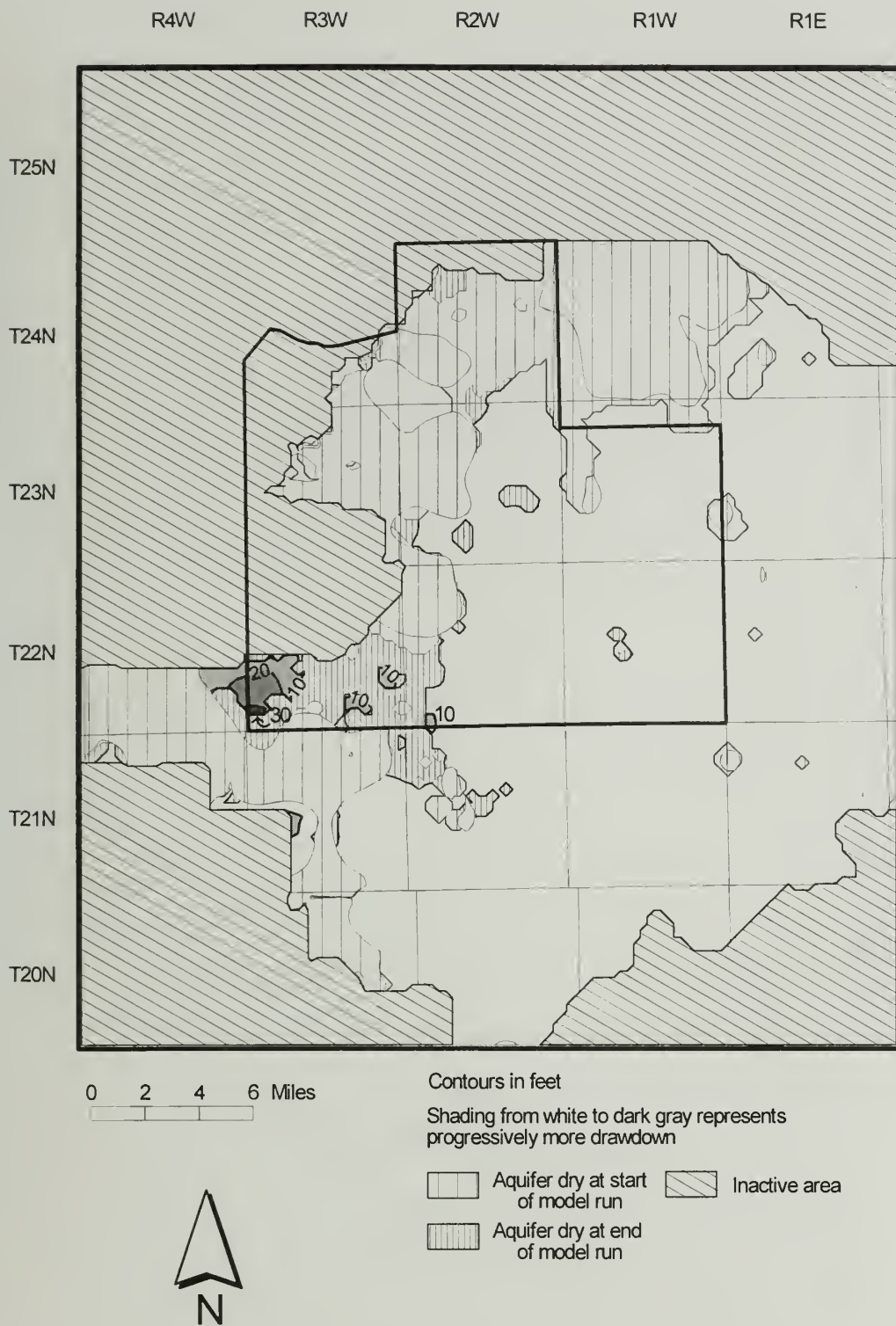


Figure 51. Drawdown in the Glasford aquifer layer in the Emden scenario (numbers indicate feet of drawdown)

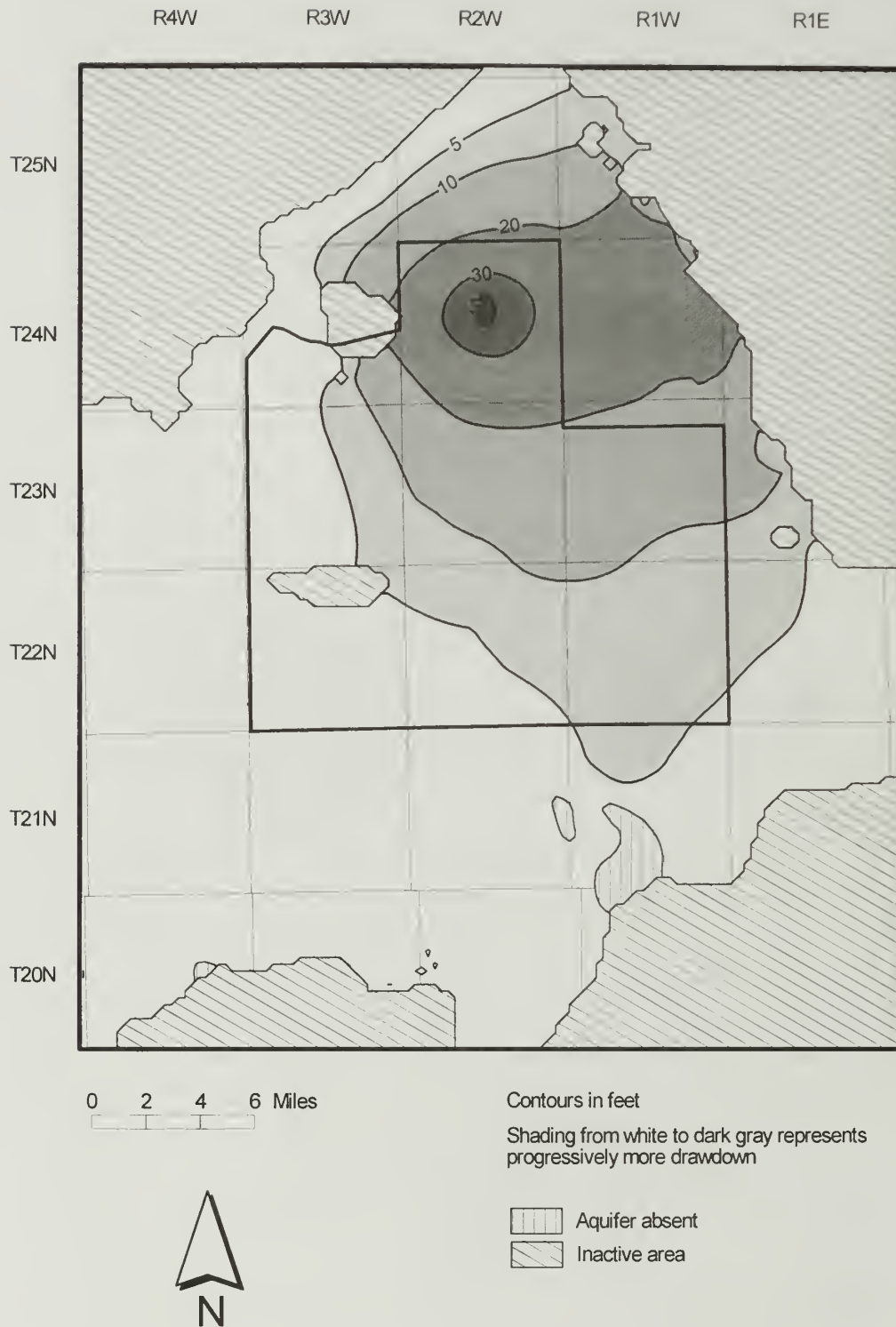


Figure 52. Drawdown in the Sankoty-Mahomet Sand aquifer layer in the Mackinaw scenario (numbers indicate feet of drawdown)

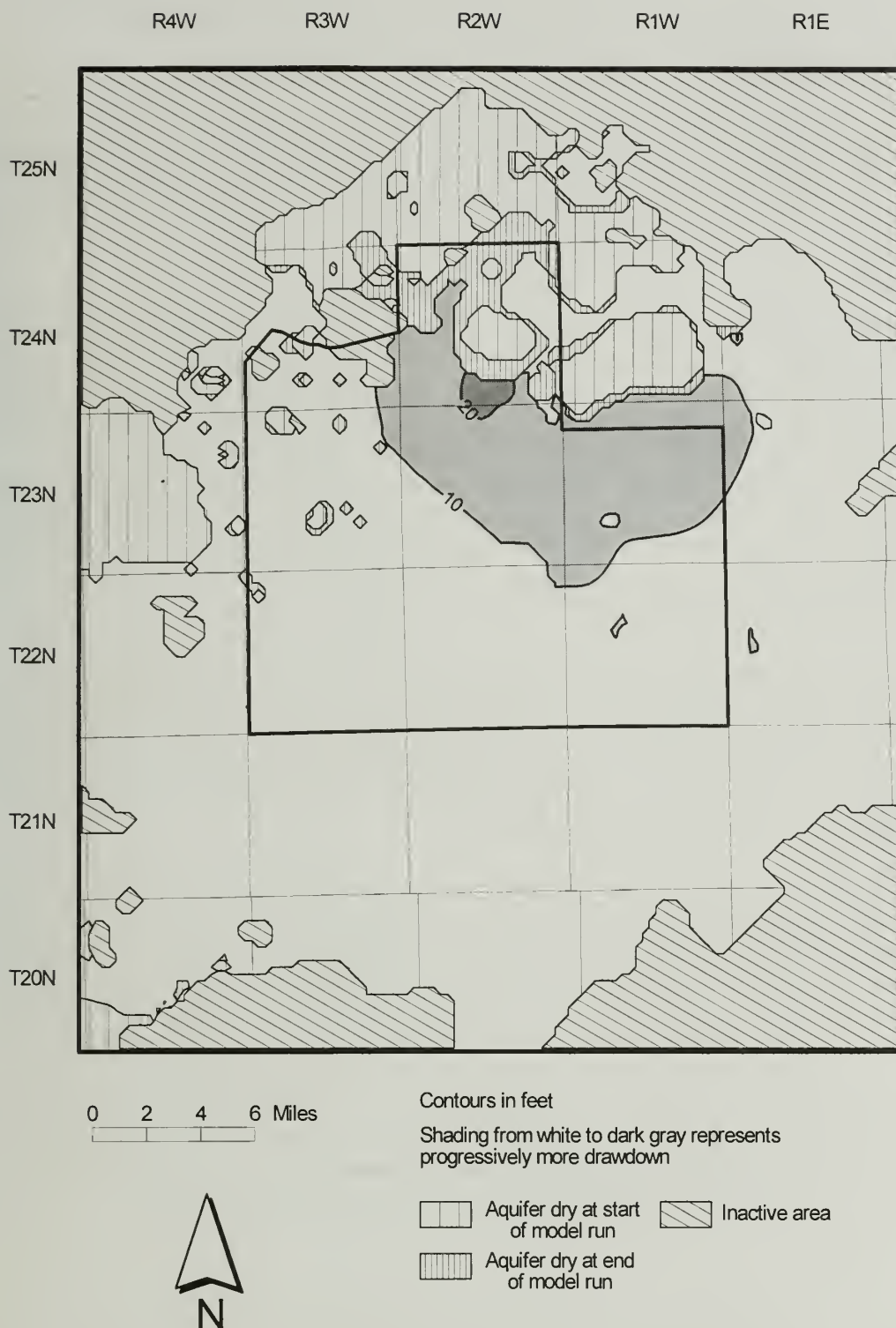


Figure 53. Drawdown in the lower Glasford/upper Banner aquifer layer in the Mackinaw scenario (numbers indicate feet of drawdown)

aquifer layer. One of the connections between the aquifers is inside the 20-foot drawdown contour depicted on figure 53. Some areas became unsaturated during the model run as a result of the lowering of water levels in the aquifer; but this should be expected adjacent to portions of the layer that were already dry.

The Glasford aquifer layer drawdown of the potentiometric surface was minimal. Figure 54 shows a small area of a 10-foot drawdown. Just south of this is an area that went dry in the model run due to the pumpage from the Sankoty-Mahomet Sand aquifer layer. In the center of this dry area, the Glasford aquifer layer is hydraulically connected to the lower Glasford/upper Banner aquifer layer.

Hopedale Scenario

The location of the wells in this scenario are in Sections 17, 18, and 19, T23N, R3W. Drawdown in the potentiometric surface of the Sankoty-Mahomet Sand aquifer from simulated pumpage at the Hopedale site is shown on figure 55. The aquifer layer is thick at this location, and there is a hydraulic connection between all three aquifer layers and the Mackinaw River. Because of this connection, the upper aquifer layers and the river recharge the Sankoty-Mahomet Sand aquifer layer as ground water is withdrawn. Drawdown in the potentiometric surface of the Sankoty-Mahomet Sand aquifer layer was minimal, about 8 feet at the well field, due to the recharged water from above. The areal extent of drawdown is also small, even though the boundary of the aquifer, the bedrock high to the northwest, is only a few miles away. Drawdown maps were omitted for the lower Glasford/upper Banner and the Glasford aquifer layers; there was less than 5 feet of drawdown in the upper two aquifer layers from pumpage at this simulated well field.

This scenario is much different from the other three because of the influence of the Mackinaw River and the two shallower aquifer layers. An additional model run was conducted to provide some assurance that the hydraulic parameters which influence the amount of surface water being induced into the Sankoty-Mahomet Sand aquifer layer were reasonable, and to determine the sensitivity of the model to changes in these parameters. When the vertical conductivity of the river bed was reduced by a factor of ten, only minimal differences were seen in the predicted drawdowns of the potentiometric surfaces.

This scenario indicates that major changes to the regional ground-water flow system in the area accounts for the reduced drawdown simulated in the potentiometric surface. In table 4, where no well-field pumpage is being simulated, the ground-water flow system provides about 1,673,380 ft³/day (12.5 mgd) of ground water to the Mackinaw River and the contribution to the ground-water flow system from the Mackinaw and Little Mackinaw Rivers is only 116,049 ft³/day (0.87 mgd). In the simulation, most of the 15 mgd withdrawn by the well field is derived by reducing the natural ground-water discharge to the rivers by 1,673,380 ft³/day (12.5 mgd) and by increasing the ground-water recharge from the rivers to 386,337 ft³/day (2.89 mgd) (table 6). The baseflow for the Mackinaw River at this location was estimated to be about 10.3 mgd (Singh and Adams, 1973) during 7-day, 10-year low-flows, which suggests that pumpage from this well field could significantly reduce or stop the flow of the Mackinaw River during periods of low flow. For reference, the average flow of the Mackinaw River near this location was approximately 80 mgd (Singh and Ramamurthy, 1991).

Impacts to Existing Wells

For each model scenario, the number of existing wells that fell within the 20-foot drawdown contour of the potentiometric surface of the Sankoty-Mahomet Sand aquifer was estimated using existing well records and information gathered from private residents during the field portion of this project. All well records were considered, including those known to be screened in one of the upper two aquifer units. The goal of this effort was to determine the possible impacts to existing wells from the pumpage at each well field. This information was included here to provide an estimate of the potential magnitude of well interference problems that could

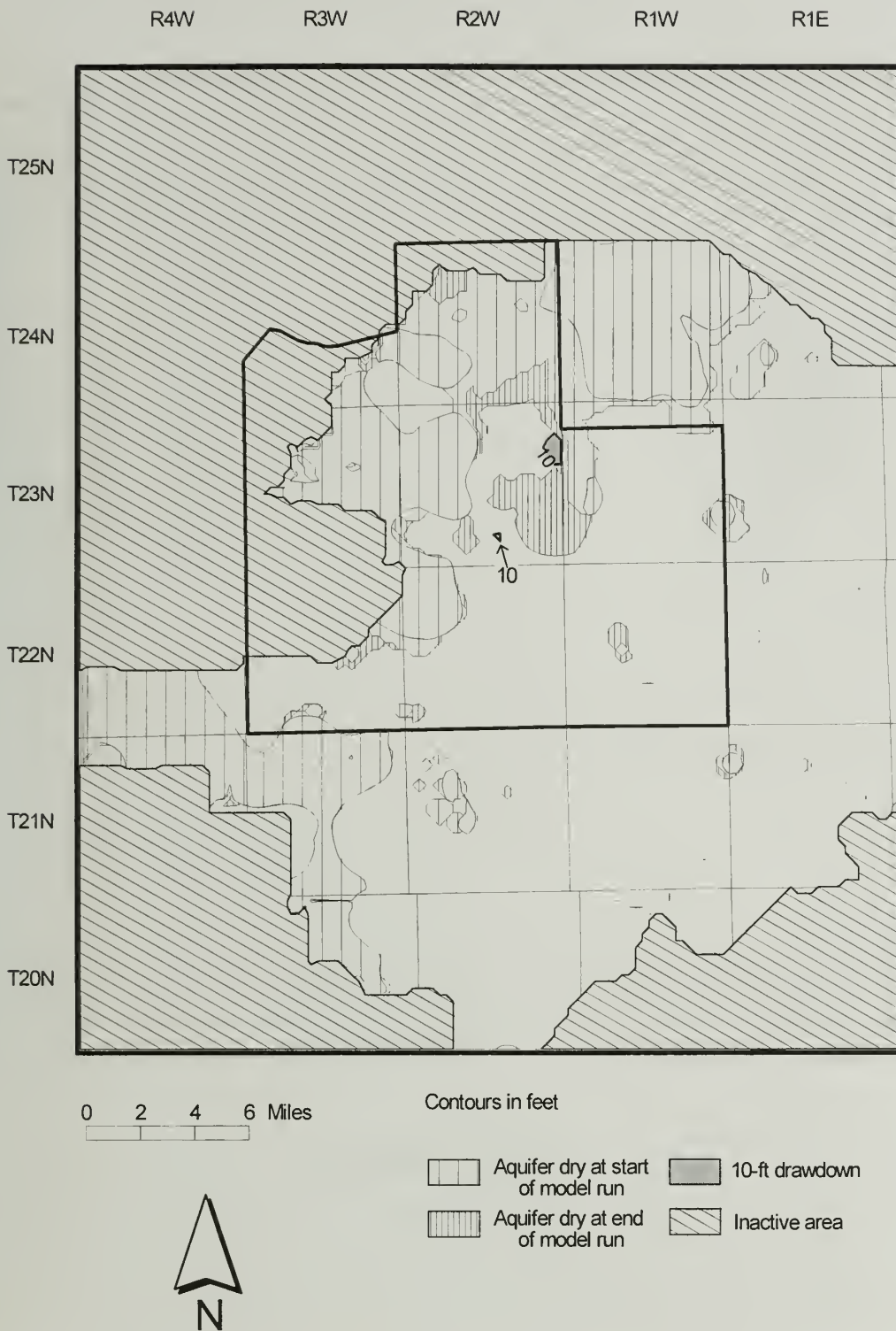


Figure 54. Drawdown in the Glasford aquifer layer in the Mackinaw scenario

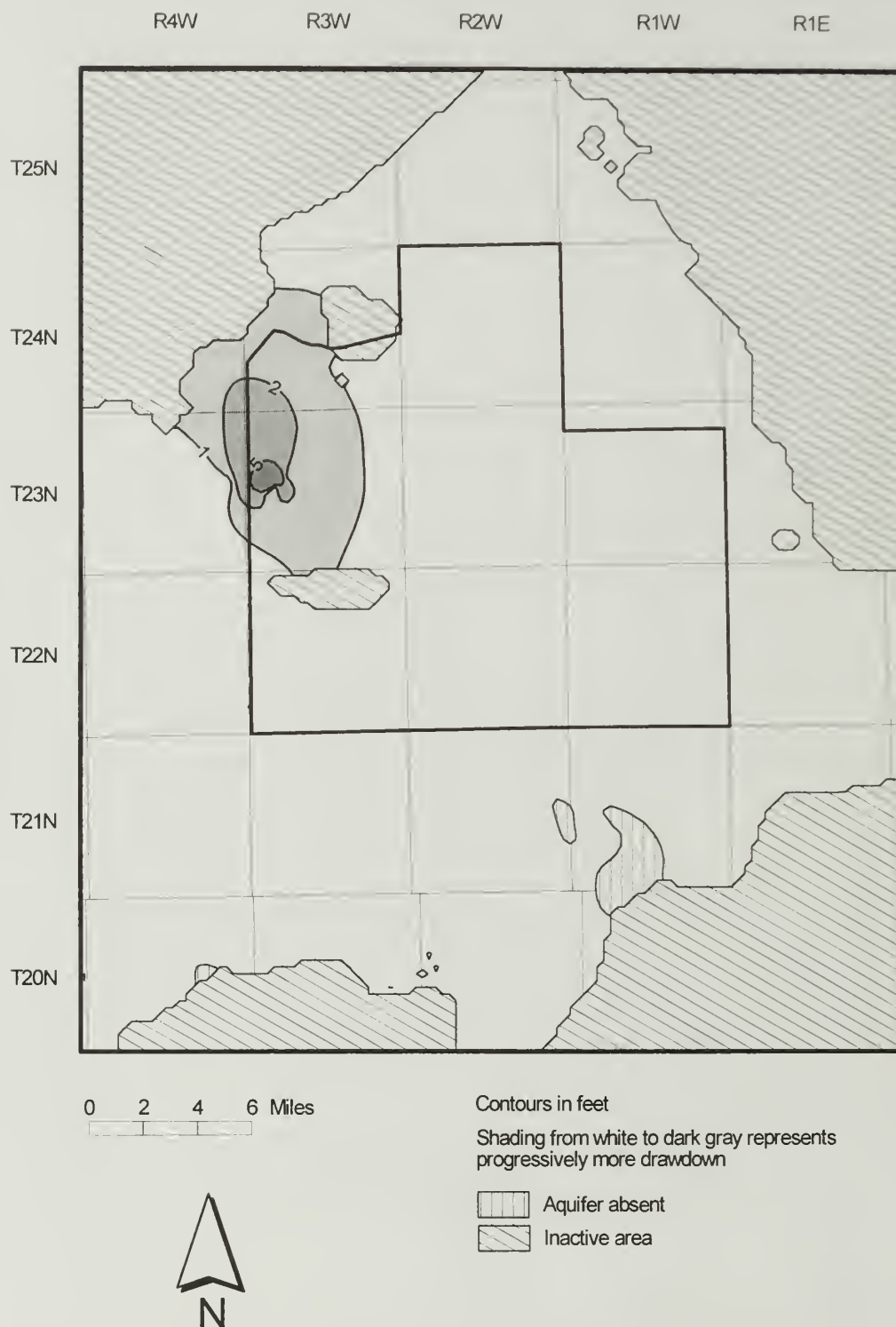


Figure 55. Drawdown in the Sankoty-Mahomet Sand aquifer layer in the Hopedale scenario (numbers indicate feet of drawdown)

Table 6. Net Ground-Water Flow Budgets of the Different Boundary Conditions Generated for the Hopedale Scenario

<i>Boundary condition</i>	<i>Inflow ft³/day (mgd)</i>	<i>Outflow ft³/day (mgd)</i>
Model edges	532,480 (3.98)	1,567,750 (11.7)
Mahomet and Kenny Bedrock Valleys	532,480 (3.98)	
Mackinaw Bedrock Valley		1,175,300 (8.79)
Western boundary		392,450 (2.94)
Streams	679,162 (5.08)	1,127,220 (8.43)
Mackinaw River	21,744 (0.16)	
Little Mackinaw River	364,593 (2.73)	
Sugar Creek above Hartsburg	292,825 (2.19)	
Sugar Creek below Hartsburg		763,223 (5.71)
Kickapoo Creek		249,380 (1.86)
Prairie Creek		114,617 (0.86)
Production wells		2,500,336 (18.7)
Infiltration from precipitation	3,985,215 (29.8)	
Net budget	5,196,857 (38.9)	5,195,306 (38.9)

Notes: ft³/day - cubic feet per day
mgd - million gallons per day

occur if a well field were actually put in place at these locations. The 20-foot contour was decided upon as the minimum drawdown that would likely require a well to be evaluated for possible impacts. An evaluation of a well does not mean that the well will necessarily be impacted; it means that a more detailed investigation of the well depth and pump setting should be conducted for that well. These data, especially the pump setting, were not readily known by most of the well owners.

Drawdown of the water level in a well can cause three impacts to occur: (1) the water level in the aquifer drops below the bottom of the well; (2) the water level drops below the pump, but water remains in the well below the pump; and (3) the water level in the well lowers, but remains above the pump. If the water level in the aquifer drops below the bottom of the well, the well is dry and unable to produce water. If this occurs, a deeper well could be drilled. If the water level drops below the pump, but not below the bottom of the well, the pump can be lowered and the well still may be able to supply the needed water. If the water level stays above the pump, the well will remain active, but there may be some reduction in the pumping capacity of the pump.

The number of existing wells to be evaluated for each scenario varied from no wells for the Hopedale scenario, where the maximum drawdown was only 8 feet, to 400 wells for the Armington scenario, where the 20-foot drawdown contour extended over about 100 square miles. For the Emden scenario, only an estimated 45 wells would need to be reviewed for possible impacts. The small number of wells at this site is due to the smaller

area impacted and the lack of subdivisions in the area. The Armington scenario would require further evaluation with approximately 250 of the 400 impacted wells being completed in the two shallower aquifers. In the Mackinaw scenario, about 300 wells would require further evaluation to determine possible impacts.

Discussion of the Scenario Results

The model runs suggest that the drawdown in the potentiometric surfaces caused by a large well field would vary significantly, depending on where it is located within the study area. The Armington scenario had the greatest predicted drawdown; it is located where part of the surrounding Sankoty-Mahomet Sand aquifer is thin. This reduces the transmissivity and causes the cone of depression to be deeper and have greater areal extent. Because of connections with the shallower aquifers, drawdowns in the shallower aquifers were significant at this location as they provided a larger portion of the ground water to the well field as compared to the Emden and Mackinaw scenarios.

In the Hopedale scenario, the majority of the pumpage was of ground water that had previously discharged naturally to the Mackinaw River. This has both positive and negative implications. Pumpage from this well field would cause virtually no adverse impacts to existing wells. However, during a drought, when the flow in the Mackinaw River may approach the low-flow reported by Singh and Adams (1973) of 10.3 mgd, this pumpage may divert a large portion of the flow in the river to the aquifer. The impacts to the Mackinaw River downstream of the well field may need to be assessed before seriously considering this location for a well field.

The shallower aquifers will be impacted by a 15 mgd well field in the Sankoty-Mahomet Sand aquifer in the study area, except along the Mackinaw River in the western part of the study area with direct hydraulic connection. This is significant because it creates a "new" system of equilibrium for the three-aquifer system and means that wells in the shallower aquifers may be impacted in some areas. Ground-water discharge to streams will be reduced. For example, the pumpage at the Armington site increased the contribution of the streams to the aquifers from 402,420 ft³/day (3.01 mgd) (table 4) to 945,224 ft³/day (7.07 mgd) (table 5), meaning there was an additional 542,984 ft³/day (4.06 mgd) of induced recharge from the streams in the model area. In addition, the natural ground-water discharge to streams was reduced from 2,807,240 ft³/day (21.0 mgd) to 2,330,541 ft³/day (17.4 mgd), or about 476,699 ft³/day (3.6 mgd). Combined, the net loss of surface water flow due to pumpage at the Armington site is 1,019,683 ft³/day (7.66 mgd).

In the Emden scenario, the predicted drawdown in the potentiometric surface of the Glasford aquifer layer was unexpected, based on the aquifer test conducted at this site. During the aquifer test no leakage was detected from the shallower aquifers into the Sankoty-Mahomet Sand aquifer. However, several reasons may explain why no drawdown occurred in the Glasford aquifer potentiometric surface during the 7-day test. The simulated well field was pumping 10,400 gpm until steady-state conditions were achieved instead of 909 gpm pumped for seven days during the aquifer test. This stressed the system significantly more. The high-capacity well installed for the aquifer test was at the same location as the northeastern most well of the ten wells at the simulated well field. This well is the farthest from the connection with the upper aquifers of the ten wells, and seven days may not have allowed sufficient time to detect leakage from the upper units. However, even if the test had been run longer, drawdown in the shallower aquifer may not have been apparent in the aquifer test data because, at the pumping rate used, the drawdown in the Sankoty-Mahomet Sand aquifer may not have been enough to induce leakage.

Evaluation of Aquifer Yield

The above described scenarios were tested to evaluate the effects of pumpage on the aquifer and existing wells. In order to evaluate the potential yield of the aquifer, greater pumpage was tested. Using the calibrated

ground-water flow model, a test run was completed that included simultaneous pumpage from three of the four scenario well fields. Specifically, the Hopedale well field was run at 15 mgd, the Emden well field was run at 7.5 mgd, and the Armington well field was run at 15 mgd, for a total withdrawal from the three well fields of 37.5 mgd. The total pumping rate was chosen somewhat arbitrarily. The purpose of this model run was to determine if the model would converge; and, if so, to evaluate the drawdown observed. If the aquifer could not sustain the withdrawal rate, the model would not have converged and the aquifer would have gone dry. The model did converge. It ran to steady-state conditions to simulate continuous, long-term pumpage. The resulting drawdown of the potentiometric surface is shown in figure 56. The model results suggest that the aquifer can maintain withdrawals of greater than this amount of pumpage if withdrawals are distributed across the model area.

As a second test, the Hopedale scenario sustained a ground-water withdrawal rate of over 37.5 mgd when run individually. This large amount of pumpage is possible because of the direct connection between the aquifers and the Mackinaw River. Much of the ground water being pumped was diverted from the Mackinaw River, the flow from which would be diminished considerably in times of drought. If the Mackinaw River were no longer able to supply recharge to the well field, then more ground water would be used by the well field, and the drawdown in the aquifers would increase.

Major Drought Simulation

The drought in 1988-1989 focused attention on the need for development of a reliable, long-term water supply. Under the climate conditions witnessed in the last 100 years, such a severe drought can be expected to occur on average once every 50 years (James Angel, Illinois State Climatologist, personal communication, 1998). However, such statistical probabilities do not preclude the recurrence of a similar drought, or even a more severe drought, in the future. It was not within the scope of this study to evaluate the probability of the occurrence of extreme climatic events, but a model run was conducted to illustrate the magnitude of drawdown that can be expected during a major drought. The importance of precipitation to the Sankoty-Mahomet Sand aquifer is underscored by the flow budget of the model, which shows infiltration of precipitation accounting for 80 percent of the water coming into the model.

There are four ways in which a major drought could affect the model area: (1) a reduction in the amount of recharge from the infiltration of precipitation, (2) a reduction in the hydraulic head of the aquifers where they leave and enter the model area, (3) the lowering of stream stage and even the drying up of some streams, and (4) the increase in pumpage from greater demands. An assumption used in the major drought simulation was that there was a very dry period in the late fall through early spring, when the ground-water system is normally recharged. This dry period would be preceded then followed by moderately dry growing seasons in which the amount of infiltration not used by plants or evaporated is insignificant. The net effect of the overall dry period and the timing of the very dry period was an 18-month window when recharge is nonexistent. This would be similar to the 1988-1989 drought experienced by the Bloomington-Normal area. During this 18-month period, the area received only 23.2 inches of precipitation in scattered events. The largest single precipitation event was only 1.56 inches.

To simulate a major drought, the model was changed from steady-state to transient mode with a two-year period starting in January of a moderately dry year. This was necessary in order to change the model parameters during the drought run. The starting conditions were taken directly from the output of the calibrated steady-state model. To account for changes in storage, a storativity of 0.0003 and a specific yield of 0.15 were assigned to the model. For the first 90 days, from January through March, conditions in the model remained the same as they were under the steady-state conditions. Starting in April, the recharge to the top layer was cut off as most of the infiltration is intercepted by plants, allowing very little water to reach the water table. The constant-head values

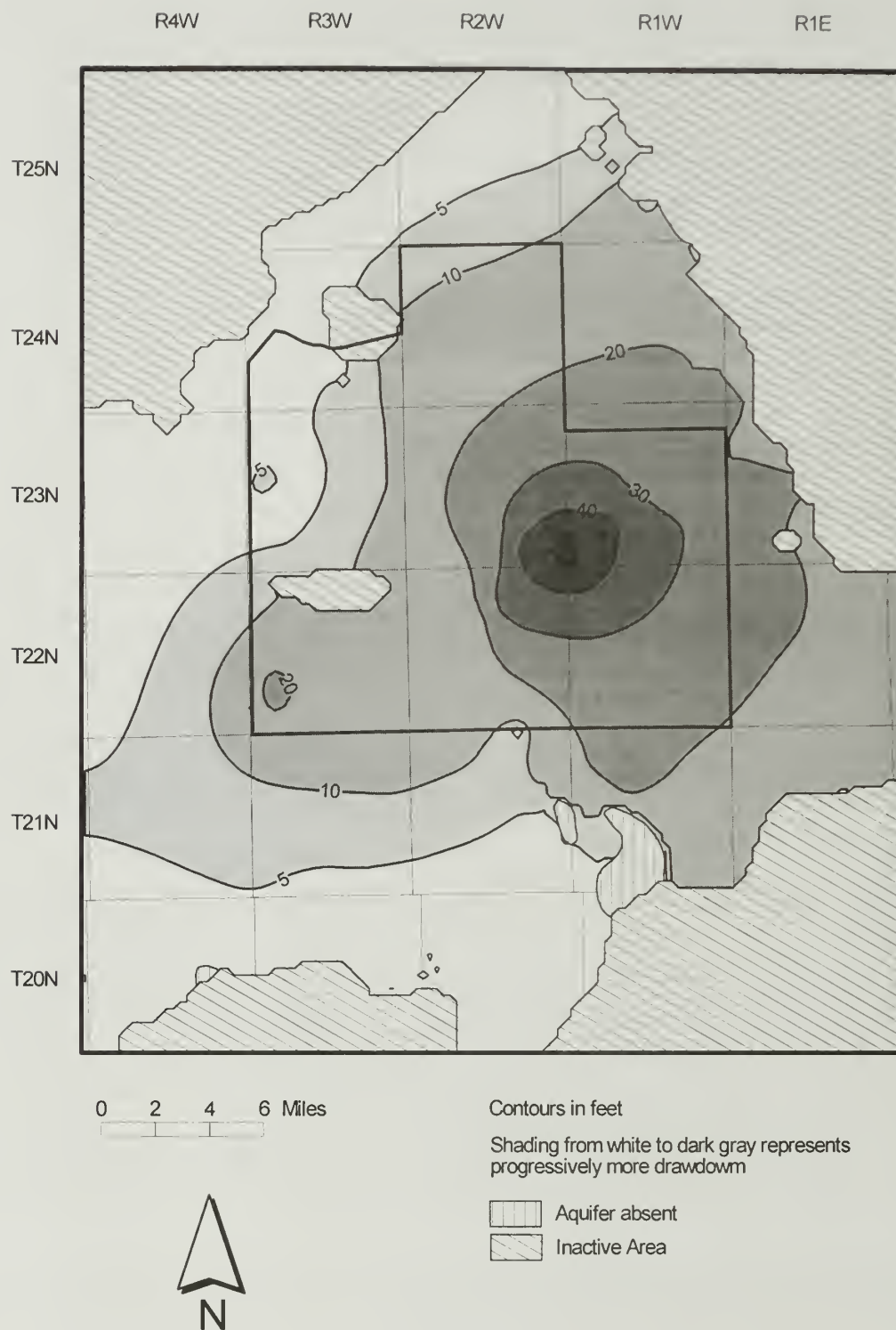


Figure 56. Drawdown in the Sankoty-Mahomet Sand aquifer layer with 3 well fields pumping together a total of 37.5 mgd (numbers indicate feet of drawdown)

along the north and west sides of the model were lowered by 5 feet to account for regional water-level declines and the effects of irrigation in Mason County.

As the simulated drought continued into the fall, the recharge in the model did not resume as it normally would when the plants stop using water. The initial conditions in the model were changed again in January of year 2 when the impact of the dry fall starts to more dramatically affect water levels. Changes included dropping all constant-head values an additional 5 feet, dropping the stage of the rivers 1 to 2 feet, eliminating infiltration in the smaller tributaries, and increasing the pumpage by 33 percent. In 1988, Normal pumped 33 percent more water than in 1994. A similar percentage was assumed for the other municipalities in the model area. The decline in head at the model edges was estimated from 1988 data at Mason County (Sanderson and Buck, 1995), and Peoria County (Burch and Kelly, 1993). Stream-flow data from 1988 (Coupe et al., 1989) showed that the Mackinaw River and Sugar Creek did not go dry, and Kickapoo Creek was dry for only three days. Therefore, the important leakage provided by Sugar Creek was maintained throughout the drought because of the upstream discharge of treated wastewater.

The decline in water level produced by the drought scenario does not appear to be significant over most of the model area (figure 57). The principal reason for this is the large amount of water that comes out of storage in the aquifer as the water levels decline. The declines are significant near locations at which the constant head conditions were lowered. This result contrasts greatly with the much larger drawdowns predicted by Clark (1994) for the 1988 drought in Mason County for which the normal ground-water levels depend on much higher rates of infiltration, and the aquifer is subject to intense irrigation.

CONCLUSIONS AND RECOMMENDATIONS

The ground-water flow system in the study area is more complex than suggested by earlier research. Understanding this complexity is the key to successful modeling; the complexity creates significant difficulties in determining the impacts on the system of further development of the ground-water resources of the area. Because of the completion of this study, model results are available that provide the basis to evaluate the potential locations for a regional well field. More importantly, this report refines the understanding of how water moves through the ground-water flow system, and reveals the cause-and-effect relationships that exist between the aquifers and a regional well field. All of the information gathered for this study—geology, water levels, water chemistry, hydraulic properties, and water use—were important components in understanding the flow system and in developing the ground-water flow model. Without all of this information, a reliable understanding of this complex flow system would not have been possible.

The potential impacts of pumpage of a regional well field on ground-water levels were found to be variable in the study area. In the four scenarios tested with the model, maximum drawdown ranged from 8 to 55 feet. In other locations, with less favorable conditions, more drawdown may occur. Researchers found that up to 400 wells may be impacted in one of the four tested scenarios. The number of wells impacted was heavily dependent on location of the well field because of the differences in drawdown that would be expected and the differences in private well density in the study area. For example, placing the well field closer to Morton would increase the potential number of impacted wells because of the increased density of private wells.

To meet the goals of this study, researchers were to determine: (1) the quantity of ground water that a regional well field could yield, and (2) the impacts to ground-water levels and existing wells that might occur from the development of a 10-15 mgd well field. These goals have been met. The Sankoty-Mahomet Sand aquifer, along with the shallow aquifers that provide recharge to it, can supply in excess of 37.5 mgd to a

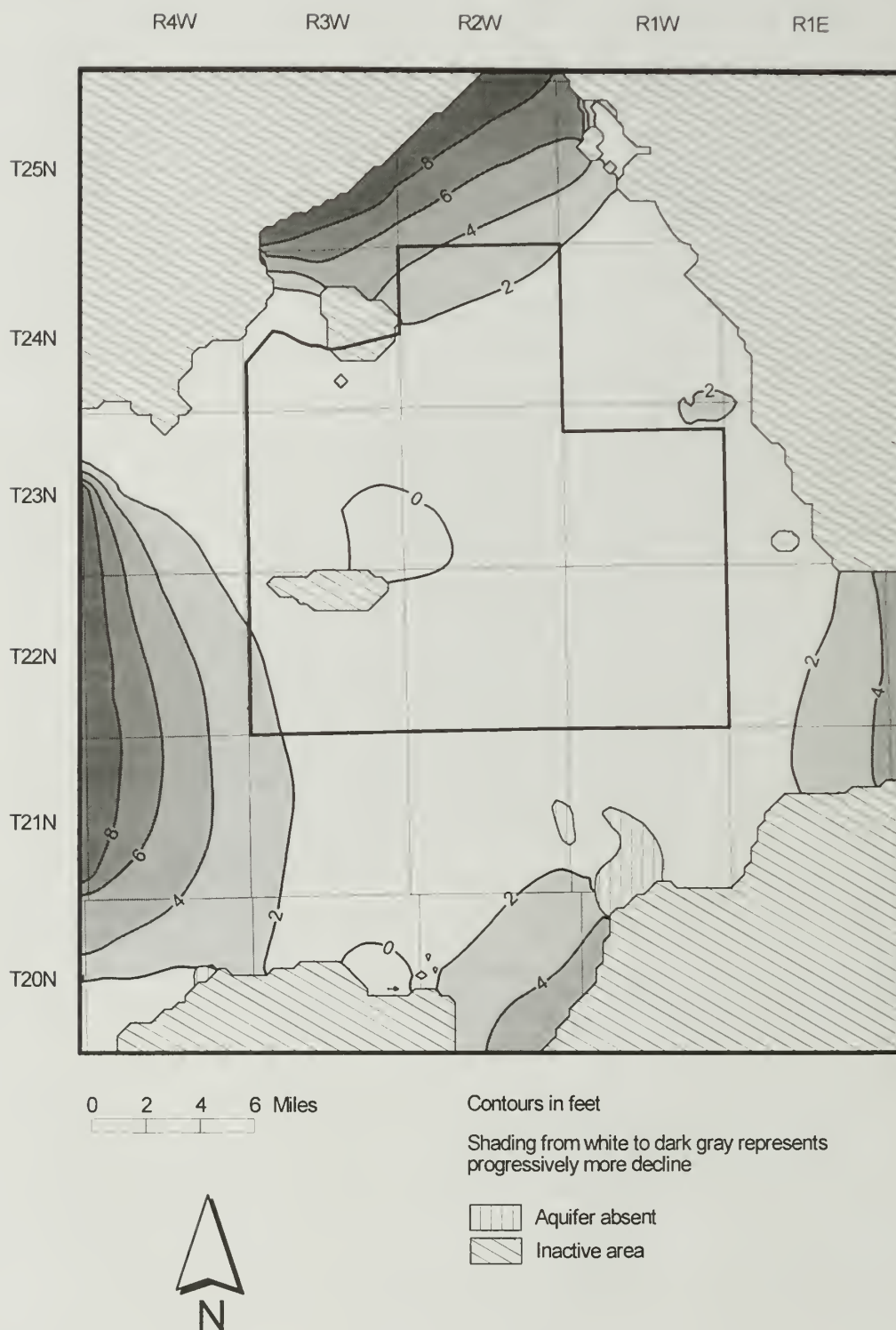


Figure 57. Declines in the potentiometric surface of the Sankoty-Mahomet Sand aquifer layer created by the drought scenario (numbers indicate feet of decline)

regional well field on a sustainable basis if the withdrawals come from areas where the Sankoty-Mahomet Sand aquifer is thick (> 100 feet) and where a nearby direct hydraulic connection exists with overlying shallow aquifers. The aquifer thickness map (figure 11), the locations of the aquifer boundaries (figure 24), and the map indicating hydraulic connections between aquifers (figure 35) together provide a good guide to the locations with the best potential for development. Placing the well field at a location with maximum aquifer thickness, with nearby connection to shallow aquifers, and where aquifer boundary effects could be minimized, would minimize the impacts from pumpage of a regional well field on water levels and on private water wells. When these factors are favorable at a location, there is a high potential for development of a regional well field.

Influences Outside the Study Area

In several of the model runs, the cone of depression reached the edge of the model. The effects of the drawdown reaching the edge are assumed to be minimal within the study area. However, drawdown beyond the edge of the model should be considered in the development of a regional water supply. Ideally, the model extent should be large enough that drawdown from simulated pumpage does not reach it. This could not be accurately determined when the study was initiated because so much was not known about the model area. It will remain a complicated task because the extent of drawdown is determined by the amount of water withdrawn, the hydrogeology, and the hydraulic properties. In other words, for a scenario pumping 5 mgd, the drawdown may not have reached the model boundaries. Furthermore, setting new boundaries may provide an adequate distance for the pumpage currently being investigated (15 mgd) to prevent the cone of depression from reaching the model boundaries, but it may not provide an adequate distance to eliminate drawdown at the model boundaries for a scenario pumping 30-40 mgd. To ensure an adequate model under any pumpage conditions, the model boundaries should be set at or beyond the aquifer boundaries. For the Sankoty-Mahomet Sand aquifer, those boundaries would include the area presented by figure 16 and the part of the Mackinaw Bedrock Valley north of the model area extending to the Illinois River.

Other municipalities (Champaign, Danville, Decatur, and Springfield) have developed or are considering development of water supplies using the Mahomet and Sankoty-Mahomet Sand aquifer. Those withdrawals have lowered or may lower the potentiometric surface and reduce lateral flow in the regional aquifer. All three of these situations could result in less water available to regional well fields. The current ground-water flow model uses boundary conditions that reflect the current status in the Sankoty-Mahomet Sand aquifer system. Therefore, if a change in pumping conditions occurs, the boundary conditions in the model might need to be refined to reflect any changes in the flow system. The better solution to this problem is to expand the model to include the entire regional aquifer system so any new stresses on the system (i.e., pumpage for a municipal supply) can be included and their effects properly evaluated.

Climate Change

Both ground-water recharge rates and water demand are dependent, in large part, on precipitation amounts. The estimates of recharge rates and water demand used in this study are based on 20th century precipitation data. If the climate were to change, precipitation, soil moisture, and water demand also could change. Although the consideration of climate change is beyond the scope of this study, climate in the 21st century could change due to natural and/or human factors (Houghton et al., 1995). The LRWPSC may wish to consider the possibility of and sensitivity to climate change when identifying and evaluating future water supply-and-demand scenarios.

Future Studies

In any dynamic system, changes will occur with time. Changes may include addition or reduction of pumpage, water-level declines or increases, water quality changes, and climate change. Public policy changes also could affect the amount of water being withdrawn, either by passage of new laws regulating ground-water use or ground-water quality, by consent of entities with an interest in ground water regarding water rights, or even by litigation over water use and water rights.

The model developed for this study incorporates all of the pertinent data currently available in the area modeled. New data will become available as wells are drilled and as other hydrogeologic data are collected in the model area. The model code itself, Visual Modflow, will undergo change and improvement in the future as ground-water science develops new uses for the model and requirements for its use. To keep the current model calibrated for future use, for instance, when water demand has increased, the model would have to be calibrated using the latest version of the computer code and new data that has become available since the last calibration of the model. To ensure the availability of the model, the current model should be updated every 5 years, and at any time a significant change in the ground-water flow system occurs, for example, when there is a large increase in ground-water withdrawals because of the addition of municipal wells.

Planning for a Regional Well Field

At some point in the future, if communities decide to move forward with plans to install a regional well field, several steps should be taken to ensure that the best location is chosen and that the selected location meets all of the required needs of the communities and the rural population. An updated model should be used to estimate ground-water level drawdown at selected sites. A detailed inventory of the existing wells in the vicinity of these selected sites should be conducted to assess the potential conflicts that might occur from pumping a well field at each selected location. When a site has been selected, an aquifer test should be conducted for a minimum of 60 days. The total pumping rate should be at least 25 percent of the anticipated capacity of the well field, higher if feasible. The results of the aquifer test could then be used to evaluate the makeup of the individual wells in the well field, either by using analytical techniques or by modifying the grid size of the model in the area of the well field to create a more dense grid in that localized area.

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GLOSSARY

- Ablation**—All processes by which snow and ice are lost from a glacier. These processes include melting, evaporation (sublimation), wind erosion, and calving.
- Aquifer**—A saturated body of earth materials that will provide a generally sustainable yield of suitable quantities of ground water.
- Aquifer Test**—A controlled field experiment to determine hydraulic properties of an aquifer using a single pumped well with a known pumping rate and at least one observation well.
- Aquitard**—A saturated geologic unit that may transmit ground water, but not in sufficient quantities to permit economic development. A source for leakage to underlying or overlying aquifers. Aquitards are significant in the study of regional ground-water flow.
- Artesian Aquifer**—An aquifer in which the water level in a well is above the top of the aquifer. The terms "leaky" and "nonleaky" are often used to describe artesian conditions. Leaky refers to the situation in which significant ground-water flow through the aquitard occurs. Nonleaky refers to the circumstances in which flow through the aquitard is insignificant.
- Cone of Depression**—A conical lowering of the water table (unconfined aquifer) or piezometric surface (in artesian confined aquifers) created by pumping a well.
- Confined Aquifer**—An aquifer that is constrained between two units of material with significantly lower hydraulic conductivity (i.e., aquitards or aquicludes).
- Confining Bed**—An aquitard or aquiclude that is contiguous to a confined aquifer.
- Crops Out**—The verb form of "outcrop". See outcrop.
- Darcy's Law**—The equation of ground-water flow through porous media, which states that the flow (Q) through a unit cross section of material (A), is proportional to the hydraulic gradient (i): $Q = KiA$, where K is the hydraulic conductivity.
- Deposition**—The laying or placing of any material; specifically, the constructive process of accumulation into beds, veins, or irregular masses.
- Dewatering**—Physical process of evacuating water from a water-table aquifer, or the lowering of ground-water levels below the top of a confined aquifer that is being pumped.
- Diamicton**—A mixture of poorly sorted sand, silt, and clay commonly referred to as till.
- Discharge Area**—The point at which water flowing through an aquifer leaves the aquifer (well, lake, river, etc.).
- Drawdown**—Difference between the nonpumping or static water level and the pumping or dynamic water level.
- Fluvial**—Produced by the action of a stream or river.

- Formation**—A body of geologic material identified by lithic characteristics and stratigraphic position; it is mappable at the earth's surface or traceable in the subsurface. The formation is the basic unit in rock stratigraphic classification.
- Geohydrologic Unit**—An aquifer, a confining unit (aquitard), or a combination of aquifers and confining units comprising a framework for a reasonably distinct hydraulic system. See Hydrostratigraphic Unit.
- Geophysical Exploration**—Use of geophysical techniques (e.g., electric, gravitational, magnetic, seismic, or radioactive) to gather information on the physical properties of the earth.
- Gradient**—The steepness of a slope.
- Ground-Water Divide**—A ridge in the water table or potentiometric surface from which the ground water represented by that surface moves away in both directions.
- Head**—The elevation to which water rises at a given point as a result of reservoir pressure; for example, the water-level elevation in a well.
- Homogeneity**—Having identical characteristics everywhere; an aquifer is homogeneous if the magnitude of all significant properties are independent of position in the aquifer; synonymous with uniformity.
- Hydraulic Conductivity**—Capacity of water-bearing material to transmit water, measured by the quantity of water passing through a unit cross section in a unit time under a unit hydraulic gradient (gallons per day per square foot).
- Hydraulic Gradient**—The rate of change of total head per unit of distance.
- Hydrograph**—A graph showing stage, flow, velocity, or other characteristics of water with respect to time. A stream hydrograph commonly shows rate of flow; a ground-water hydrograph commonly shows water level or head.
- Hydrostratigraphic Unit**—A geologic unit or combination of units that exhibit hydraulic continuity over a significant area. See Geohydrologic Unit.
- Lacustrine**—Pertaining to, produced by, or formed in a lake or lakes, e.g., lacustrine silts deposited on the bottom of a lake. Fine-textured, water-laid deposits.
- Lithification**—Conversion of a newly deposited, unconsolidated sediment into a coherent, solid rock, involving processes such as cementation, compaction, desiccation, and crystalization.
- Mass Water-Level Measurement**—Measurement of the depth-to-water in a large number of wells over a short period of time (one to two weeks) to provide a "snap-shot" of the potentiometric surface of an aquifer across a given area at that specific instant in time.
- Mean Absolute Error**—The mean of the absolute values of the error measurements.
- Member**—A lithostratigraphic unit of subordinate rank, comprising some specially developed part of a formation.

- Model**—A representation of a physical system. Ground-water models may be physical, electric analog, or mathematical.
- Observation Well**—Any well used to measure water levels or obtain water samples.
- Outcrop**—Geologic material that appears at the ground surface. For example, the sand-and-gravel outcrop is just east of Mackinaw.
- Outwash**—Stratified glacial material (chiefly sand and gravel) removed or "washed out" from a glacier by meltwater streams and deposited in front of or beyond the end moraine or margin of an active glacier.
- Permeability**—Capacity of a porous rock, sediment, or soil for transmitting a fluid; a measure of the relative ease of fluid flow under unequal pressure.
- Potable**—Safe and drinkable for human use.
- Potentiometric Surface**—An imaginary surface representing the total head of ground water and defined by the level to which water will rise in a well.
- Pumping Level**—Water level in a well being pumped at which the observed water level has little or no change within a reasonable period of time.
- Radius of Influence**—Distance from the discharge well to the edge of the cone of depression.
- Recharge**—Water added to an aquifer from infiltration of precipitation or movement to the aquifer from surrounding hydrogeologic units.
- Recharge Area**—Localized or regional area at which water enters an aquifer.
- Scour**—Erosion by the action of air, ice, or water; the removal of surface material by rock fragments dragged along by a glacier.
- Specific Capacity**—Rate of discharge of water from a well divided by the drawdown of water level within the well; varies with duration of discharge, which should be stated if known. If the specific capacity is constant except for the time variation, it is roughly proportional to the rate at which water flows through the aquifer. The relationship between discharge and drawdown is affected by the construction of the well, its development, the character of the screen or casing perforation, and the velocity and length of flow up the casing.
- Specific Yield**—Ratio of the volume of water that the rock or soil, after being saturated, will yield by gravity to the volume of rock or soil. The definition implies that gravity drainage is complete. In the natural environment, specific yield generally is observed as the change that occurs in the amount of water in storage per unit area of unconfined aquifer as the result of a unit change in head. Such a change in storage is produced by the draining or filling of pore space and, therefore, is dependent on particle size, rate of change of the water table, time, and other variables. Hence, specific yield is only an approximate measure of the relationship between storage and head in unconfined aquifers. It is equal to porosity minus specific retention (water held by soil particles).

- Static Water Level**—Water level in a nonpumping well outside the area of influence of any pumping well. This level registers one point on the water table in a water-table well or one point on the piezometric surface in a well intersecting an artesian aquifer.
- Storage Coefficient**—Volume of water that an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. In a confined system, water derived from storage with decline in head comes from the expansion of the water and compression of the aquifer; similarly, water added to storage with a rise in head is accommodated partly by compression of the water and partly by expansion of the aquifer. In an unconfined system, the amount of water derived from or added to the aquifer by these processes generally is negligible compared to that involved in gravity drainage or filling of pores; hence, in an unconfined system the storage coefficient is virtually equal to the specific yield.
- Stratigraphic Unit**—A body of adjacent geologic materials recognized as a unit in the classification of rock sediments.
- Thalweg**—A line connecting the lowest or deepest points along a streambed or valley.
- Till**—Unsorted and unstratified drift deposited directly by and underneath a glacier without subsequent reworking by meltwater; consists of a heterogeneous mixture of clay, silt, sand, gravel, and boulders ranging widely in size and shape.
- Transmissivity**—Formerly called the coefficient of transmissibility. Rate at which water flows through a unit width of the aquifer (perpendicular to flow) under a unit hydraulic gradient; product of hydraulic conductivity and saturated thickness.
- Unconfined Aquifer**—Also called a water-table aquifer; an aquifer in which the water table forms the upper boundary.
- Water Table**—That surface in an unconfined body of water at which the pressure is atmospheric; defined by levels at which water stands in wells that penetrate the body of water just far enough to hold standing water.
- Well Loss**—Component of drawdown in a well due to frictional losses from turbulent flow as water passes through the screen or well face and inside the casing to the pump intake.

APPENDIX A. AQUIFER CHARACTERIZATION STUDY

This section summarizes the major points previously published in the aquifer characterization study by Herzog et al. (1995a and b), which described the nature and occurrence of the sand-and-gravel aquifers located within southwestern McLean and southeastern Tazewell Counties. The data gathered for the aquifer characterization study helped develop an improved understanding of the hydrogeology and ground-water flow system in the study area. Some of the interpretations made in the aquifer characterization study have since been updated and improved upon because the area investigated was expanded, new data were incorporated, and the ground-water flow model provided further insight into the continuity of the hydrogeologic units.

To complete the aquifer characterization (Herzog et al., 1995a and b), data from Survey files gathered prior to 1993 were augmented with data from surface geophysical surveys and test drilling, mass measurements of water levels, aquifer tests, water quality samples, raingages, and stream measurements collected for the aquifer characterization portion of this study. Surface geophysical surveys were conducted along 45 miles of highway rights-of-way using seismic refraction. Data from these surveys were used to improve upon earlier maps of bedrock topography and to guide selection of sites for test drilling. Test holes were drilled through the entire sequence of glacial deposits and into bedrock at 25 locations. An observation well was installed in the Sankoty-Mahomet Sand aquifer at 22 of the test-hole sites. At 6 of the 22 sites, at which a second significant aquifer was encountered at a shallower depth, a second observation well was installed in a separate test hole. No wells were installed at three of the sites because the aquifer was too thin for installation of a well, or absent altogether. Water levels in the observation wells were measured biweekly to provide information on the long-term and seasonal fluctuations in water-level elevations. Data from two mass water-level measurements were used to create potentiometric-surface maps for the aquifers. Water samples were collected from the observation wells to gather regional water quality information and isotope data, both of which were used to provide information on ground-water mixing characteristics and recharge.

The test-hole and seismic-refraction data from the aquifer characterization study provided several new interpretations of the geology. These included:

- Evidence of hills and depressions in the bedrock surface, greater in size and number than previously mapped. The most significant of these bedrock hills, located south of Hopedale (Tazewell County), protrudes through the Sankoty-Mahomet Sand aquifer over several square miles, so no aquifer exists inside this area.
- Thick, clay-rich, fine-grained lacustrine deposits instead of the sand-and-gravel aquifer between three bedrock hills in the center of the study area and at several locations in the south-central and eastern part of the study area. These deposits limit the thickness and areal extent of the Sankoty-Mahomet Sand aquifer at these locations.

The bedrock hills and lacustrine deposits effectively divide the Sankoty-Mahomet sand aquifer into four reaches, each containing significant areas in which the aquifer is more than 100 feet thick. The aquifer is greater than 150 feet thick east of Mackinaw and between Tremont and Hopedale. West of Hopedale, it is directly overlain by shallower sand-and-gravel aquifers for a combined thickness of more than 150 feet. Although the Sankoty-Mahomet Sand aquifer is thinner in parts of the study area than previously thought, it also is much thicker in others. Consequently, the estimated volume of the Sankoty-Mahomet Sand aquifer in the confluence area was not significantly different from previous volume estimates, even though its geometry and hydrogeology are more complex than previous work indicated. When the aquifer is thin or absent, the potential of developing a high-capacity well field is limited.

Kempton et al. (1991) noted that the total thickness of the sand-and-gravel deposits above the Sankoty-Mahomet Sand aquifer and in the Banner and Glasford Formation within the confluence area were less than 20 feet, an inadequate thickness for high-capacity wells. Available data were insufficient to map them separately, so the aquifer characterization study (Herzog et al., 1995a) followed the example of Kempton et al. (1991) and mapped them as one hydrogeologic unit, informally called the upper Banner-Glasford Formation aquifers.

Results from two aquifer tests confirmed the interpretation of the hydrogeologic data collected during the aquifer characterization study, that the Sankoty-Mahomet Sand aquifer is variable in thickness and areal extent. Analysis of aquifer test data collected in the southwest corner of the study area shows transmissivity values in the Sankoty-Mahomet Sand aquifer are high, ranging from 264,000 to 298,000 gpd/ft. Storage coefficient values of 1.0×10^{-4} to 8.9×10^{-5} are typical of a confined aquifer. Aquifer test results indicate the aquifer is prolific and confirm that the boundary created by the bedrock hill south of Hopedale would affect drawdown in local wells and could influence the potential yield of the aquifer in that area. Transmissivity values derived from an aquifer test at Mackinaw are even greater, ranging between 337,000 to 354,000 gpd/ft; storage coefficient values of 4.9×10^{-4} to 9.8×10^{-4} also are typical of a confined aquifer. The aquifer test results show that the shallow sand-and-gravel aquifers appear to be a source of leakage into the deeper Sankoty-Mahomet Sand aquifer.

Herzog et al. (1995a) indicated that local precipitation is the main source of recharge to the aquifer system, supplemented by lateral ground-water flow into the study area. The occurrence and distribution of precipitation varies seasonally and from year to year, a characteristic of the continental climate. Spring precipitation typically is more widespread than summer precipitation, which tends to be mostly from thunderstorms or large storm cells associated with frontal systems. Most of the annual precipitation in the study area falls during the spring and summer months.

In the western part of the study area near the Mackinaw River, fluctuations of water levels measured in observation wells completed in the Sankoty-Mahomet Sand aquifer parallel the changes in the river stage. In this area, the Sankoty-Mahomet Sand aquifer occurs near land surface or is hydraulically connected to the river where the aquifer directly underlies a shallower aquifer. A ground-water divide in the Sankoty-Mahomet Sand aquifer trends roughly southeast to northwest across the study area. Ground water north of the divide flows northward out of the confluence area into the Mackinaw Bedrock Valley portion of the aquifer and toward the Illinois River, the regional discharge area for this part of the Sankoty-Mahomet Sand aquifer (Marino and Schicht, 1969). South of the divide, ground water flows toward the west and the Illinois River, the regional discharge area for this part of the aquifer (Walker et al., 1965).

Estimates by Herzog et al. (1995a) of the ground-water recharge rate differed little from previous estimates (Walton, 1965; Kempton and Visocky, 1992; and Wilson et al., 1994). Aquifer-test data demonstrated that the Sankoty-Mahomet Sand aquifer has an excellent ability to transmit water, and a sustained water supply of 10-15 mgd on a long-term basis was thought to be possible.

APPENDIX B. HYDROGEOLOGIC DATABASE

API#	P#	Lambert x	Lambert y	Twp	Range	Section	Quarters	LSE	Total depth	Depth	Top elev	Shallow sand			Bottom elev (real)	First marker	
												Log thickness	Min thickness	Bottom elev (max)		Depth	Elev
121792290600	287286	3047767	2709935	23N	2W	4	NW SE SE	630	137								
121792161500	176815	3038103	2706333	23N	2W	7	SE SE NE	630	122								
121790069900	176816	3037910	2706294	23N	2W	7	SW SE NE	634	134							45	589
121792233300	176817	3036691	2704452	23N	2W	7	NE SW SE	600	107							38	597
121792272000	267667	3035301	2708410	23N	2W	7	NE NE NW	635	162								
121792161600	176896	3034656	2705736	23N	2W	7	NW NE SW	595	131								
121792301400	289221	3033303	2706379	23N	2W	7	SW SW NW	590	93								
121790007200	176813	3038023	2705802	23N	2W	7	NE NESW?	612	108	10	602	15	15	587	587	50	562
121792107100	176822	3039973	2707157	23N	2W	8	NW SE NW	635	117								
121790007300	176824	3046528	2705693	23N	2W	9	NW NW SE	645	135								
121790026600	176823	3046481	2708600	23N	2W	9	NW NW NE	645	159							65	580
121792300600	294655	3051306	2703787	23N	2W	10	SE SE SW	642	286							45	597
121792280300	294650	3064840	2704224	23N	2W	12	SE SE SE	658	335	6	652	6	6	646	646	55	603
121792207300	176836	3061411	2703509	23N	2W	13	NE NE NW	657	168								
121792207301	176835	3061411	2703509	23N	2W	13	NE NE NW	657	203								
1217900064200	3059288	2699005	23N	2W	14	SE SE SE	640	62		8	632	14	14	618	618		
121790026200	176837	3058297	2698803	23N	2W	14	NW SE SE	645	195								
121792211200	250319	3056434	2698883	23N	2W	14	SE SE SW	645	140	8	637	12	12	625	625	41	604
121790007500	176839	3052280	2699176	23N	2W	15	NW SW SE	645	168								
121792132600	176844	3047994	2698753	23N	2W	16	SW SE SE	640	85								
121792280600	269182	3038723	2703183	23N	2W	17	NW NW NW	630	130								
121792280700	291537	3037072	2702158	23N	2W	18	NE NE NE	640	130	37	603	6	6	597	597		
121790007600	176852	3037798	2698390	23N	2W	18	SW SE SE	630	147							55	575
121792279400	294637	3037229	2695701	23N	2W	19	SW SE NE	641	278	40	601	1	1	600	600	58	583
121792211300	185513	3041384	2695292	23N	2W	20	NW NW SE	648	157								
121792280800	269181	3055381	2693654	23N	2W	23	SE SW SW	636	110	9	627	12	12	615	615		
121792161700	176864	3064685	2695172	23N	2W	24	SE NE SE	660	109								
121792279500	294638	3053504	2688363	23N	2W	27	SW SE SE	630	305	7	623	8	8	615	615	46	584
121790007800	176870	3053153	2689338	23N	2W	27	center SE	630	106								
121792286400	284629	3050870	2688294	23N	2W	27	SW SE SW	635	106							56	579
121792244300	244283	3033360	2690547	23N	2W	30	SW SW NW	669	194								
121792211400	185529	3039943	2686632	23N	2W	32	SE NE NE	661	270							55	606
121792279600	294639	3040214	2682594	23N	2W	32	SW SE SW	648	318	6	642	19	19	623	623	31	617
121790007900	176886	3045470	2687094	23N	2W	33	SW NE NW	642	101	20	622	2	2	620	620		
MTH-27		3062390	2688307	23N	2W	36	NW NW NE	644	295	8	636	5	5	631	631	46	598

API#	Glasford sands					Second marker		Lower Glasford/upper Banner sands						
	Depth	Top elev	Log thickness	Min thickness	Bottom elev (max)	Bottom elev (real)	Depth	Elev	Depth	Top elev	Log thickness	Min thickness	Bottom elev (max)	Bottom elev (real)
121792290600	54	576	16	16	560	560			110	520	24	24	496	496
121792161500	59	571	2	2	569	569			110	520	>12	12.5	507.5	
121790069900			0	0					129	505	>5	5.5	499.5	
121792233300			0	0					81	519	>26	26.5	492.5	
121792272000	69	566	10	10	556	556			112	523	>50	50.5	472.5	
121792161600			0	0					79	516	>52	52.5	463.5	
121792301400			0	0					73	517	>20	20.5	496.5	
121790007200									93	519	>15	15.5	503.5	
121792107100			0	0					102	533	>15	15.5	517.5	
121790007300			0	0					133	512	>2	2.5	509.5	
121790026600			0	0					120	525	>39	39.5	485.5	
121792300600	67	575	34	34	541	541	109	533			0	0		
121792280300	75	583	37	37	546	546	125	533	117	541	14	14	527	527
121792207300									128	529	12	12	517	517
121792207301														
121790064200														
121790026200	75	570	2	2	568	568					0	0		
121792211200	64	581	19	19	562	562			90	555	>36	36.5	518.5	
121790007500	70	575	12	12	563	563			110	535	28	28	507	507
121792132600	70	570	>15	15.5	554.5									
121792280600			0	0					92	538	32	32	506	506
121792280700									93	547	35	35	512	512
121790007600			0	0					83	547	>60	60.5	486.5	
121792279400	67	574	12	12	562	562	79	562	81	560	52	52	508	508
121792211300			0	0					86	562	>71	71.5	490.5	
121792280800			0	0					78	558	>32		525.5	
121792161700	84	576	21	21	555	555			105	555	>4		550.5	
121792279500	72	558	7	7	551	551	96	534	92	538	7	7	531	531
121790007800			0	0					94	536	>12	12.5	523.5	523.5
121792286400			0	0					95	540	>11	11.5	528.5	
121792244300	102	567	18	18	549	549			120	549	62	62	487	487
121792211400	95	566	15	15	551	551			160	501	19	19	482	482
121792279600	69	579	2	2	577	577	101	547	93	555	58	58	497	497
121790007900			0	0					80	562	>21	21.5	540.5	
MTH-27	65	579	10	10	569	569	133	511	120	524	13	13	511	511

APPENDIX B. CONTINUED

API#	Sankoly-Mahomet Sand						Bedrock		10-acre plot
	Depth	Top elev	Log thickness	Min thickness	Bottom elev (max)	Bottom elev (real)	Depth	Elev	
121792290600						405.3			2B
121792161500									1E
121790069900									2E
121792233300									3B
121792272000									5H
121792161600									6D
121792301400									8E
121790007200									?
121792107100						383.3			6F
121790007300									4D
121790026600									4H
121792300600	160	482	121	121	361	361	281	361	5A
121792280300	167	491	150	150	341	341	317	341	1A
121792207300	160	497	>8		488.5				5H
121792207301	168	489	>35		453.5				5H
121790064200									1A
121790026200	170	475	>25	25.5	449.5				2B
121792211200									5A
121790007500	158	487	>10		476.5				4B
121792132600									2A
121792280600									8H
121792280700									1H
121790007600									2A
121792279400	222	419	40	40	379	379	262	379	2
121792211300									4D
121792280800									7A
121792161700									1C
121792279500	202	428	15	15	413	413	289	341	2A
121790007800									2B
121792286400									6A
121792244300	182	487	>12	12.5	474.5				8E
121792211400	248	413	>22	22.5	390.5				1G
121792279600	175	473	126	126	347	347	301	347	6A
121790007900									6G
MTH-27	153	491	125	125	366	366	280	364	4H

APPENDIX B. CONCLUDED

Notes:

API# - Unique identification number assigned to a well record in the ISGS database
 Bottom elev (max) - Maximum (or highest) allowable elevation for the bottom of a hydrogeologic unit; used when other data are not available, in feet
 Bottom elev (real) - Actual elevation for the bottom of a hydrogeologic unit based on an interpretation of the well log, in feet
 Depth - The depth from land surface of the top of a geologic unit, in feet
 Elev - Elevation, in feet above mean sea level
 Lambert - Refers to the Lambert Conformal Conic Projection for Illinois, a coordinate system for describing the east-west and north-south position of a point from a common reference point
 Lambert x - East-west Lambert coordinate, in feet
 Lambert y - North-south Lambert coordinate, in feet
 LSE - Land surface elevation, in feet above mean sea level
 Log thickness - Thickness of a hydrogeologic unit as interpreted from a well log, in feet
 Min thickness - Minimum allowable thickness of a hydrogeologic unit; used when other data are not available, in feet
 P# - Unique identification number assigned to a well record in the ISWS database
 10-acre plot - ISWS 2-digit description of the location of a well within a section
 Top elev - Elevation of the top of a hydrogeologic unit, in feet above mean sea level
 Total depth - Total depth of a well from land surface, in feet
 Twp - Township

APPENDIX C. LOG OF TEST HOLE MTH-27

Location: Tazewell County, Section 36.4h T23N, R02W

Elevation: 644'

Total Depth: 295'

Drilled: 6/4/97

LITHOLOGIC LOG

Depth (ft)	Unit/Description	Depth (ft)	Unit/Description
0-5	Clay, black (10YR 2/1), very silty to clayey silt; very soft; crumbly to slightly cohesive; smooth; highly organic, leached	31-33	Clay, dark grayish brown (2.5Y 4/2), moderately silty, slightly to moderately sandy with very fine to very coarse sand, very slightly pebbly with fine to medium gravel; slightly to moderately soft; moderately cohesive and calcareous; thin layer of very pebbly to clayey gravel at 35'
5-8	Clay, yellowish brown (10YR 5/4), as above but has small, slightly calcareous spots (limestone fragments?)	33-35	Sand and gravel, very fine sand to medium gravel, mostly medium sand to fine gravel, poorly sorted; overall gray color
8-10	Sand and gravel, very fine sand to fine gravel, mostly very fine to fine sand, slightly well sorted; overall yellowish brown color	35-40	Clay, grayish brown (10YR 5/2), moderately to very silty, very slightly sandy with very fine to fine sand; slightly soft to firm; moderately cohesive; moderately to very calcareous
10-14	Clay, yellowish brown (10YR 5/4), very silty to clayey silt; moderately sandy with very fine to coarse sand; soft; crumbly to slightly cohesive, leached		
	Wedron Group		Glasford Formation - diamicton
14-15	Clay, olive brown (2.5Y 4/2); very silty to silty clay; very slightly sandy with sparse very fine sand; soft; smooth; slightly to moderately cohesive; leached but with sparse coarse to very coarse sand size, sub-rounded, limestone fragments	40-46	Clay, dark gray (10YR 4/1), moderately to very silty; slightly to moderately sandy with very fine to very coarse sand, mostly very fine to fine sand; very slightly pebbly with fine to medium gravel, mostly carbonates and shale; slightly to moderately soft, moderately cohesive, slightly calcareous
15-30	Clay, gray (10YR 5/1), silty to very silty; slightly sandy with very fine to coarse sand, mostly very fine to fine sand; very slightly pebbly with fine to medium gravel size, sub-rounded to sub-angular, dolomite fragments; moderately cohesive, moderately calcareous; sparse black flecks throughout; very pebbly at 29' with fine to coarse gravel, mostly carbonates, shale, and chert? fragments	46-48	Clay, light olive brown (2.5Y 5/3); moderately to very silty; slightly sandy with very fine to fine sand; slightly soft to firm; moderately to very cohesive; moderately calcareous
30-31	Sand and gravel; very fine sand to coarse gravel, mostly very fine to medium sand, poorly to slightly well sorted; overall gray color	48-55	Clay, reddish brown? (5YR 4/3), moderately silty, moderately sandy with very fine to very coarse sand, mostly very fine to medium sand; slightly soft and cohesive; slightly calcareous; thin pebbly layer at about 52'; rock at 55' - lithic type of rock lost in cuttings

Depth (ft)	Unit/Description
55-57	Clay, grayish brown (2.5Y 5/2), moderately to very silty, slightly to moderately sandy with very fine to medium sand; slightly soft and cohesive; moderately calcareous
57-62	Clay, gray (2.5Y 5/1), as above 55'-57'
62-65	Clay, gray (2.5Y 6/1), moderately silty, slightly to moderately sandy and thin sandier layers with very fine to medium sand; slightly to moderately pebbly in thin layers with fine to medium gravel that's mostly carbonate and shale grains; slightly soft and cohesive; moderately to very calcareous
65-69	Sand and gravel, very fine sand to medium gravel, poorly sorted
69-86	Clay, gray (2.5Y 5/1), moderately to very silty, moderately sandy with very fine to medium sand; slightly pebbly with fine to medium gravel that's mostly carbonate, shale, and quartz grains; slightly soft and cohesive; thin layers of sandy gravel at 71'-72', 76'-77', and 84'-85'; gravel is mostly shale and carbonate fragments
86-87	Clay, very dark gray (10YR 3/1), moderately to very silty, moderately to very soft; slightly to moderately cohesive; moderately calcareous
87-91	Clay, as above 69'-86'
91-97	Sand and gravel, very fine sand to fine gravel, mostly medium to very coarse sand, poorly sorted; moderately silty and clayey interstitially; some thin layers of silty clay to clayey silt, gray (10YR 6/1), soft, moderately calcareous; more silty and clayey below 95'
97-120	Clay, gray (2.5Y 5/1), moderately to very silty, slightly sandy and moderately to very sandy in layers with very fine to fine sand; slightly pebbly and moderately pebbly in layers with fine to medium gravel; moderately soft; slightly cohesive; slightly to moderately calcareous; thin layers of sandy gravel at 107'-108'
120-131	Sand, very fine to very coarse, mostly

Depth (ft)	Unit/Description
	very fine to fine, moderately well sorted; mostly subrounded to rounded quartz grains with carbonate and shale
131-133	Sand and gravel, very fine sand to medium gravel, poorly sorted

Banner Formation

133-146	Clay, gray (2.5Y 5/1), moderately to very silty, slightly sandy with very fine to fine sand, very slightly pebbly with fine gravel; slightly soft and cohesive; very calcareous; small fragments of wood abundant at 146'
146-151	Clay, dark gray (2.5Y 4/1), moderately to very silty, slightly to moderately sandy with very fine to fine sand, slightly pebbly with fine gravel that's mostly carbonates and shale with quartz; slightly soft to firm, slightly cohesive, moderately to very calcareous; much sandier below 150'
151-153	Clay, grayish brown (10YR 5/2), very silty, very sandy and pebbly to a silty, clayey sand and gravel; very fine sand to medium gravel; crumbly to slightly cohesive; slightly calcareous; becomes sandier and more pebbly below 158'

Sankoty-Mahomet Sand

153-163	Sand and gravel, very fine sand to medium gravel, poorly sorted; moderately to very silty and clayey interstitially; crumbly; slightly calcareous; less silty and clayey with depth
163-175	Sand, very fine to very coarse, mostly very fine to fine, slightly to moderately well sorted
175-181	Sand and gravel, very fine sand to fine gravel, mostly very fine to medium sand, slightly well sorted
181-255	Sand, very fine to very coarse, mostly very fine to fine, moderately well sorted; interval contains thin zones with sparse fine gravel; maybe somewhat coarser with more gravel at 205'-212' and 238'-247'

Depth (ft)	Unit/Description	DRILLER'S LOG - MTH-27	
255-257	Limestone, light olive brown (2.5Y 5/3); dense; very calcareous	Depth (ft)	Description
257-275	Sand, as above 181'-255'; overall olive brown color with abundant black grains and less abundant red grains	0-9	yellow clay
275-280	Sand and gravel, very fine sand to medium gravel, mostly fine to coarse sand, poorly to slightly well sorted	9-11	sand and gravel
		11-29	very hard gray clay
		29-30	rocks
		30-31	gravel
		31-45	rocky gray clay with some gravel streaks
		45-65	green clay with gravel streaks
		65-69	gravel
		69-81	gravel with clay seams
		81-125	gray clay
		125-151	nice #12 slot sand
		151-162	wood very sandy clay
		162-170	#10 - #12 sand
		170-180	#12 - #15 sand with 10% 1/8 inch
		180-220	#10 sand
		220-237	#8 - #10 sand
		237-245	#12 - #15 sand with stones
		245-255½	#15 - #20 sand & 20% 1/8 inch
		255½-257½	limestone boulder
		257½-280	#15 - #20 Sankoty type sand (lot of reds)
		280-281	boulder
		281-295	light gray shale

Pennsylvanian bedrock

280-281	Limestone?, white (5Y 8/1); very calcareous; hard; this may or may not be in place
281-293	Shale, light gray (10YR 7/1) to gray (10YR 6/1); very stiff drilling; cuttings are smooth, moderately to very soft, slightly cohesive, noncalcareous; cuttings hydrate in the drilling fluid
293-295	Shale, dark gray (5Y 4/1) to grayish brown (2.5Y 5/2); very stiff drilling; cuttings are smooth, moderately to very soft, slightly cohesive, noncalcareous; cuttings hydrate in the drilling fluid

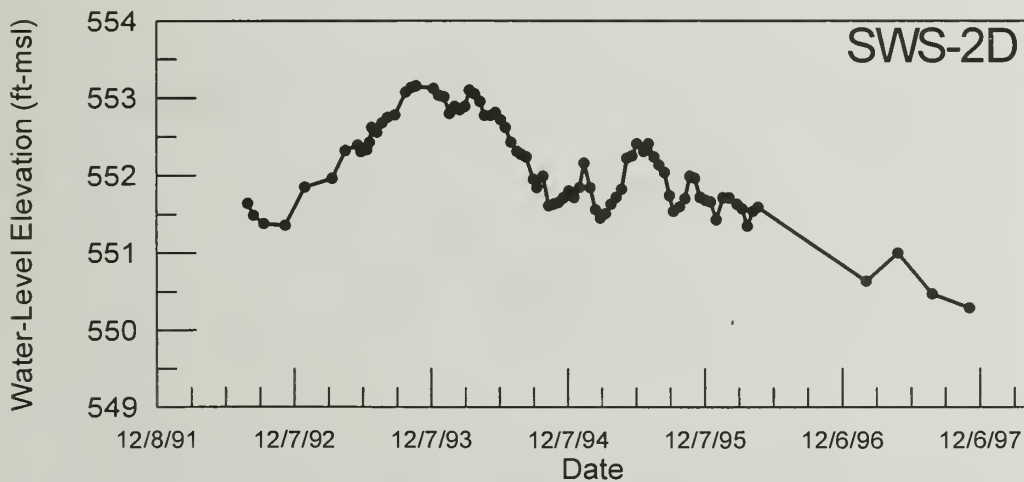
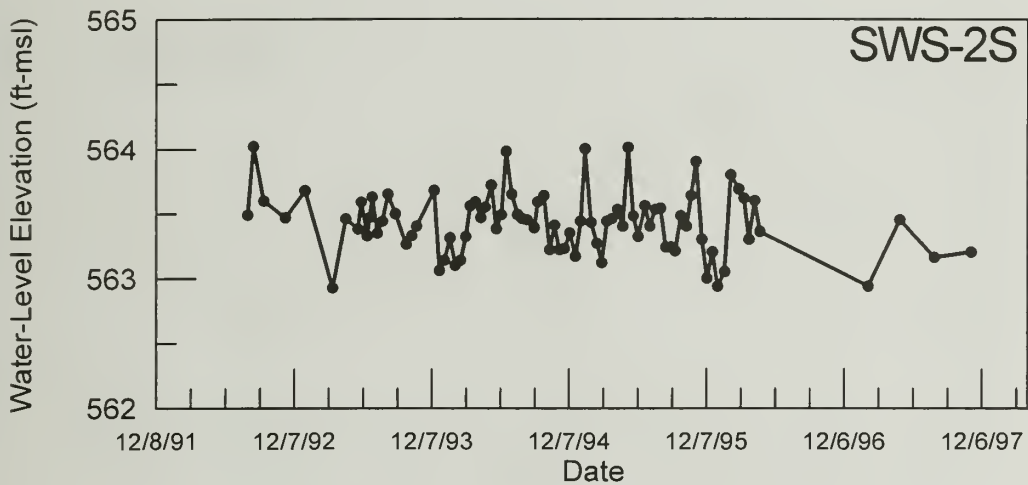
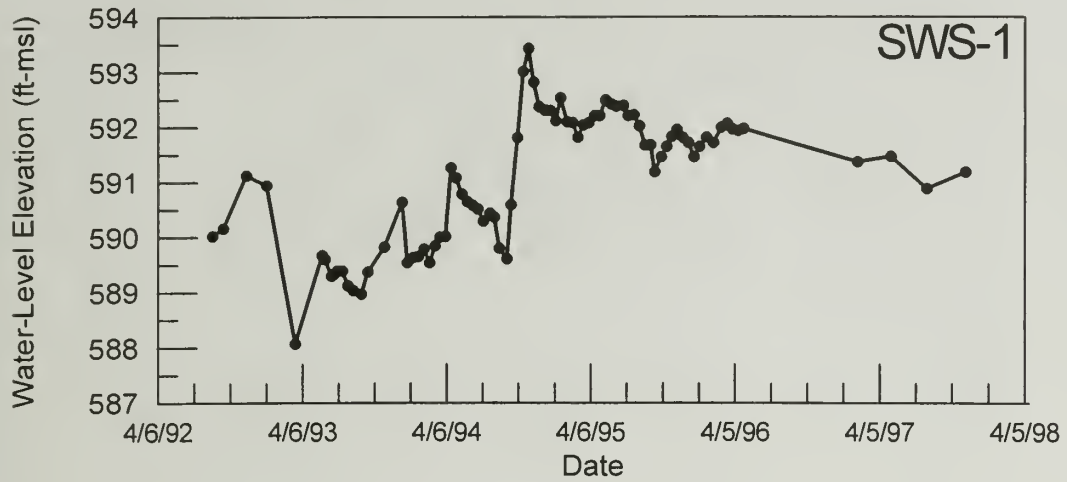
OBSERVATION WELL CONSTRUCTION: MTH-27

Set 10' of #10 slot 2" PVC screen and 230' of 2" PVC schedule 40 pipe; backfilled test hole with pea gravel to bottom of screen, backflushed with water until discharge cleared, and sand packed screen. Filled annulus with pea gravel to depth of 207' and sealed annulus with 20 pounds of bentonite chips. Continued filling annulus with pea gravel to 23' with 2 annular seals in clay units of 20 pounds of bentonite chips. Filled annulus with bentonite chips to land surface; set protective casing. Depth to bottom of screen is 237' below land surface. Surveyed elevation of the measuring point of the well (top of the casing) is 646.8 ft above mean sea level.

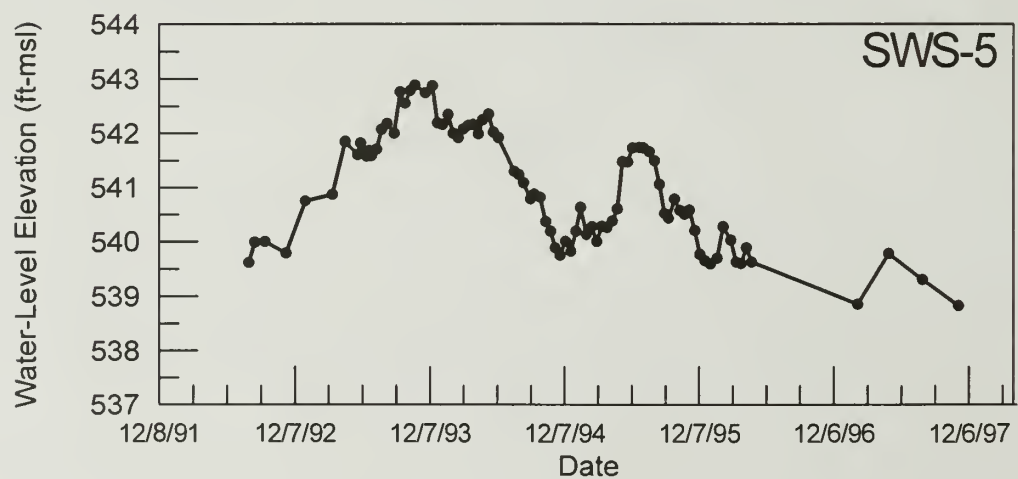
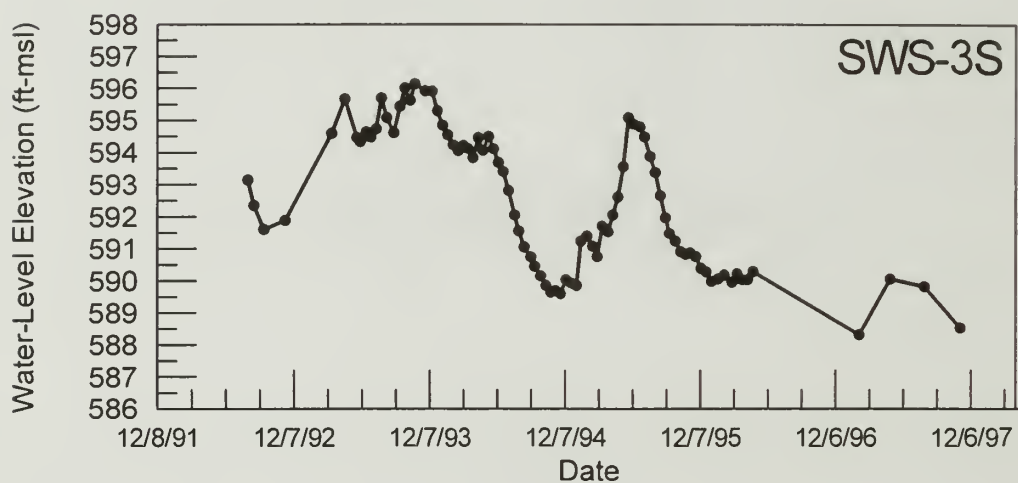
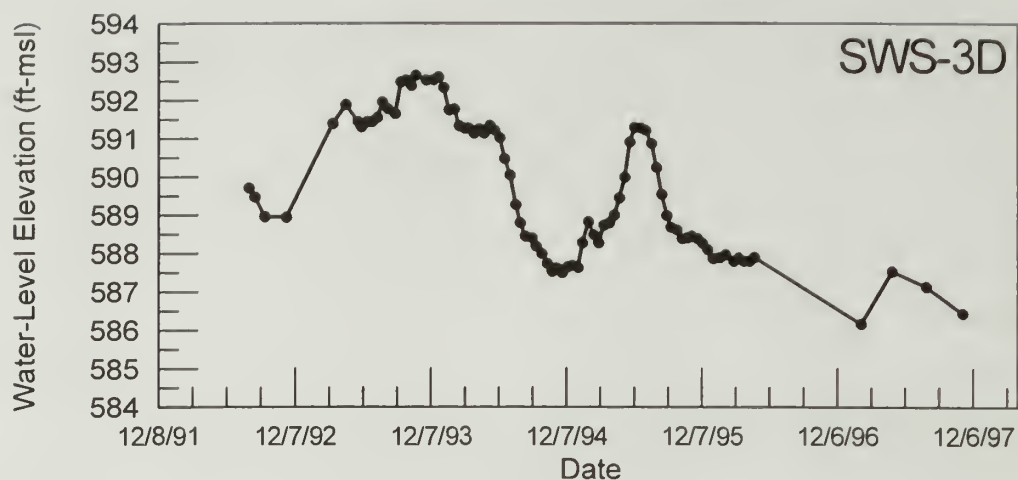
Note:

In the materials description, 2.5YR 5/1, for example, refers to the Munsell field classification. The Munsell system is a standardized method used to describe colors by comparing the color of a sample of sediment to the standard color chips in the Munsell Soil Color Charts. The Munsell notation shows hue, value, and chroma. Hue indicates a color's relation to the color. Value indicates the lightness of the color. Chroma indicates the strength of the color. R = red, Y = yellow, G = green, B = blue, and P = purple.

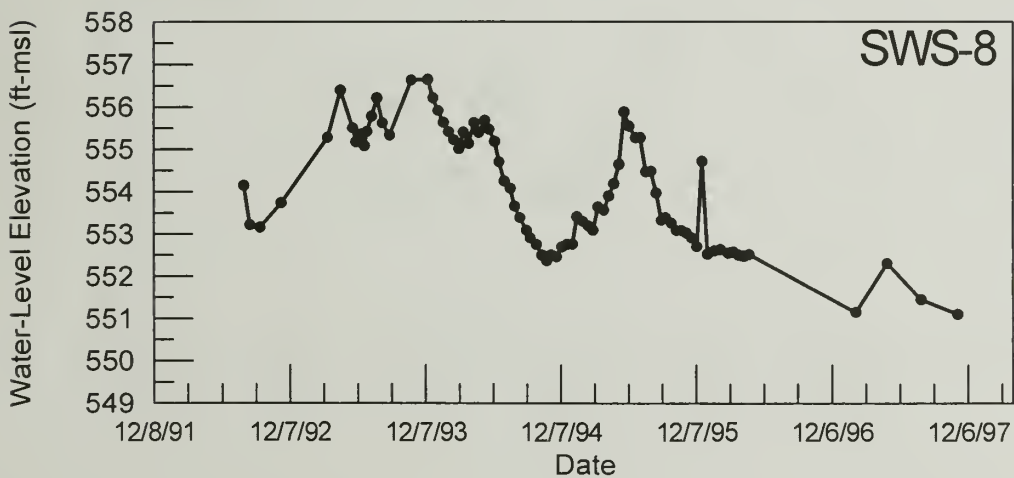
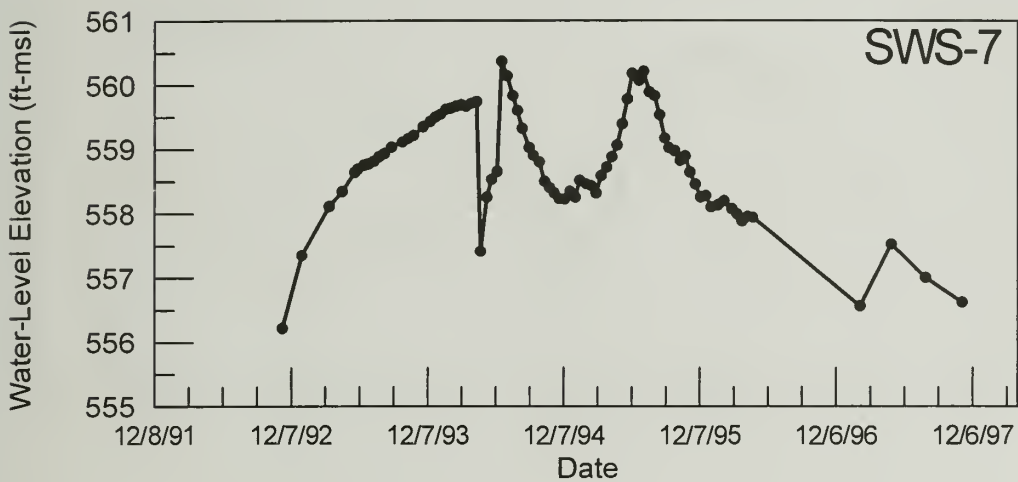
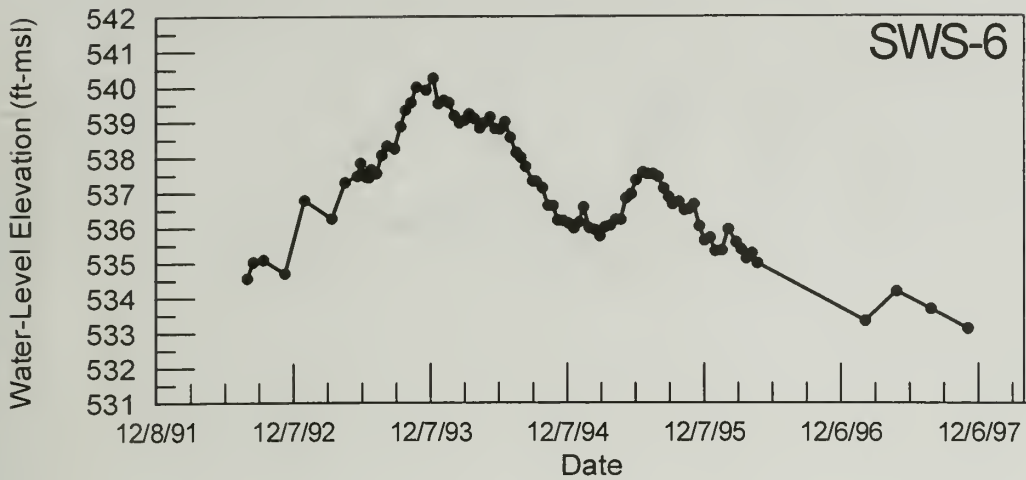
APPENDIX D. OBSERVATION WELL HYDROGRAPHS AND RIVER STAGE



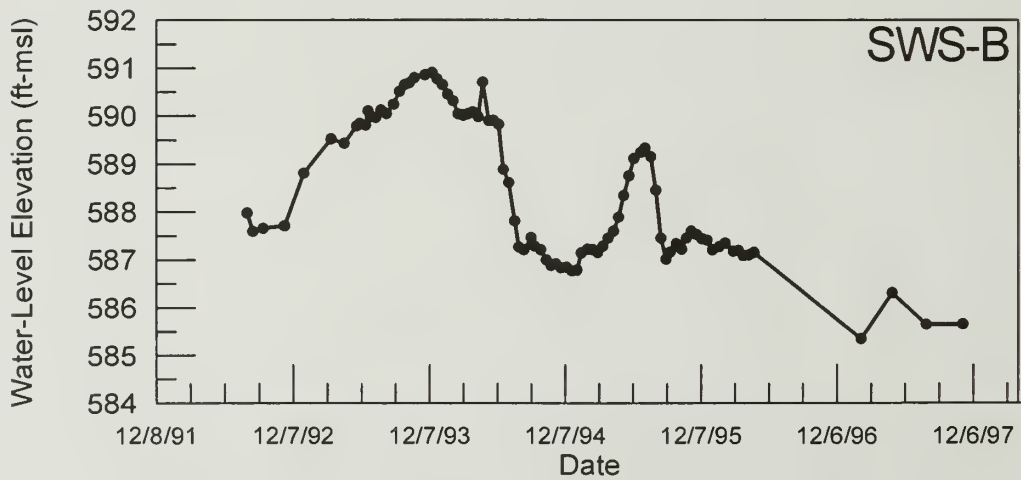
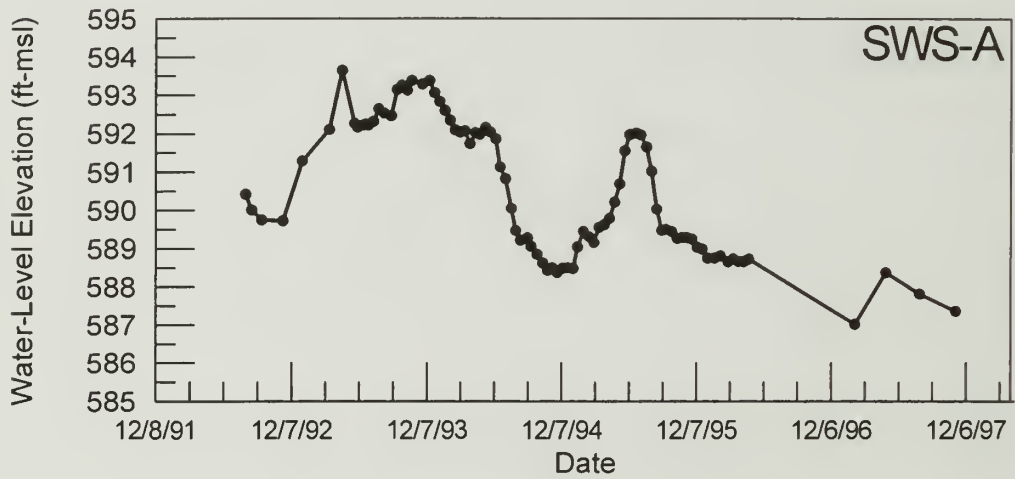
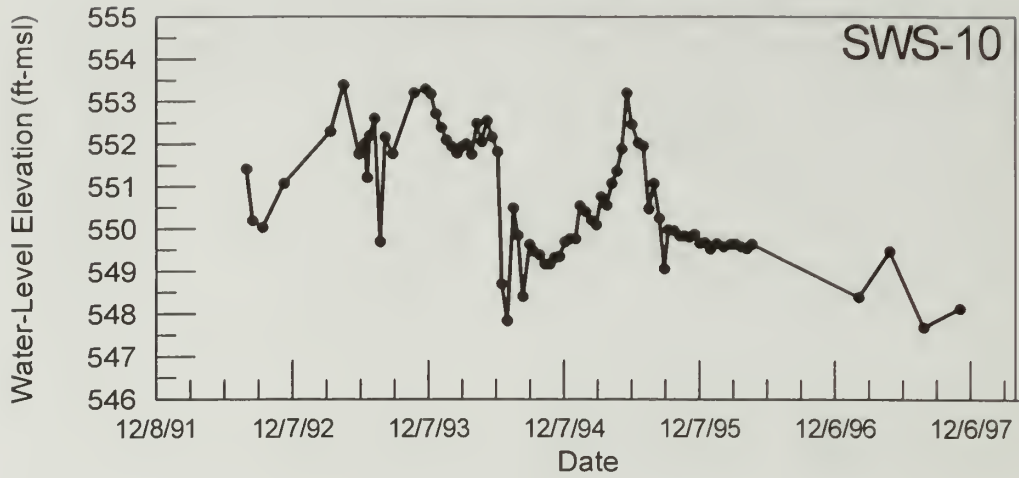
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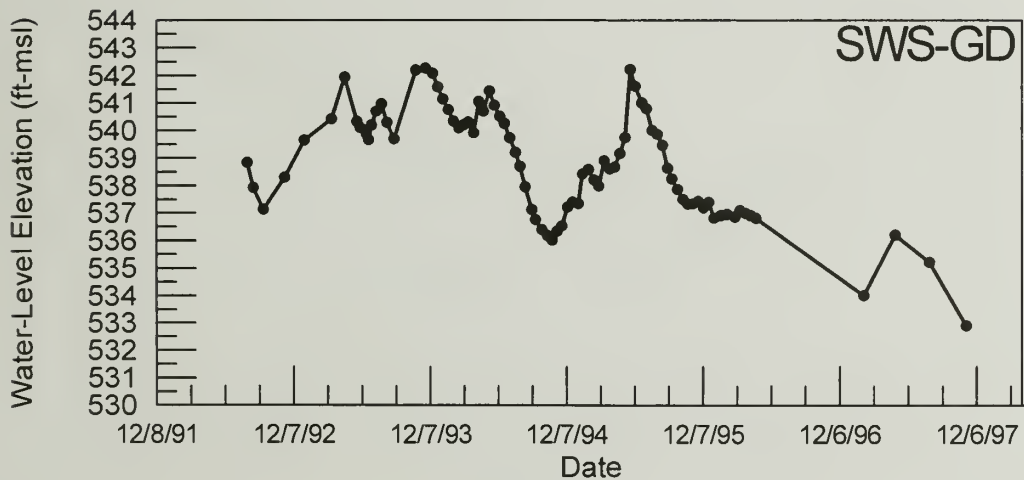
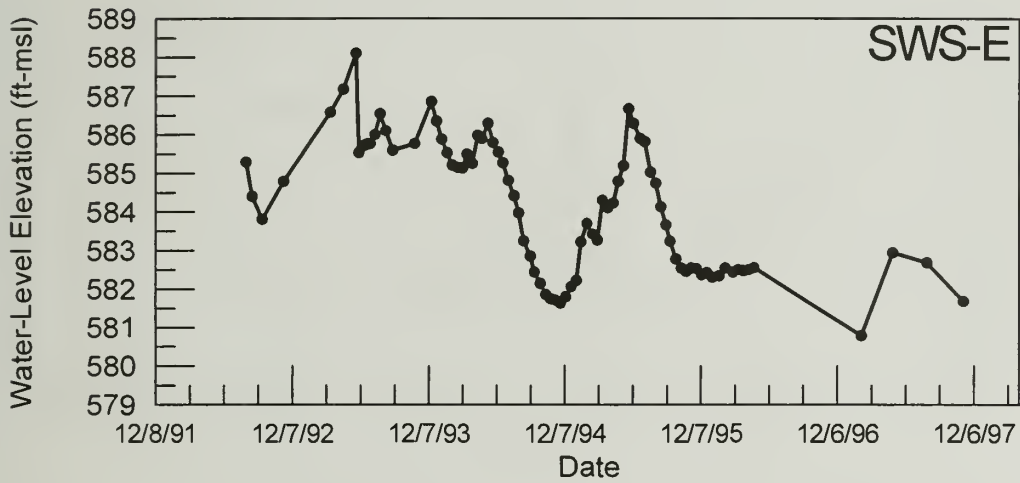
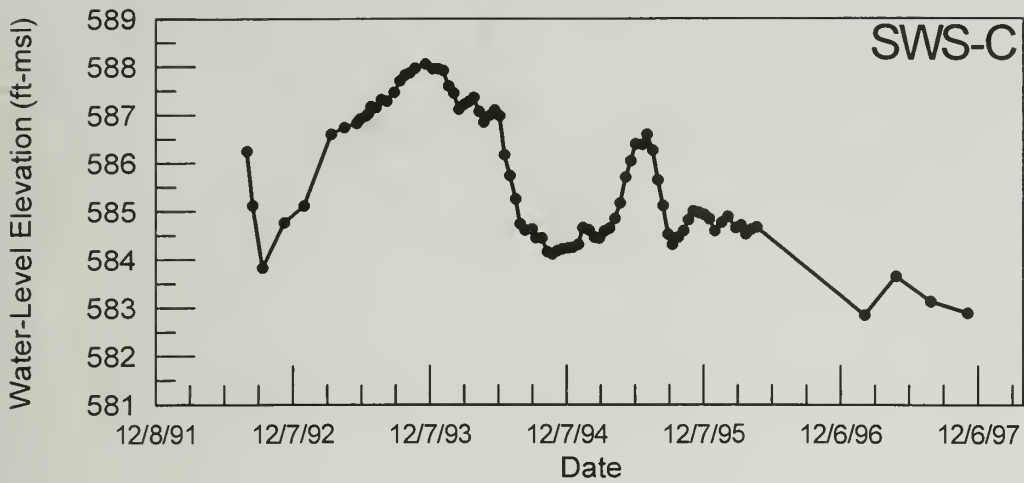
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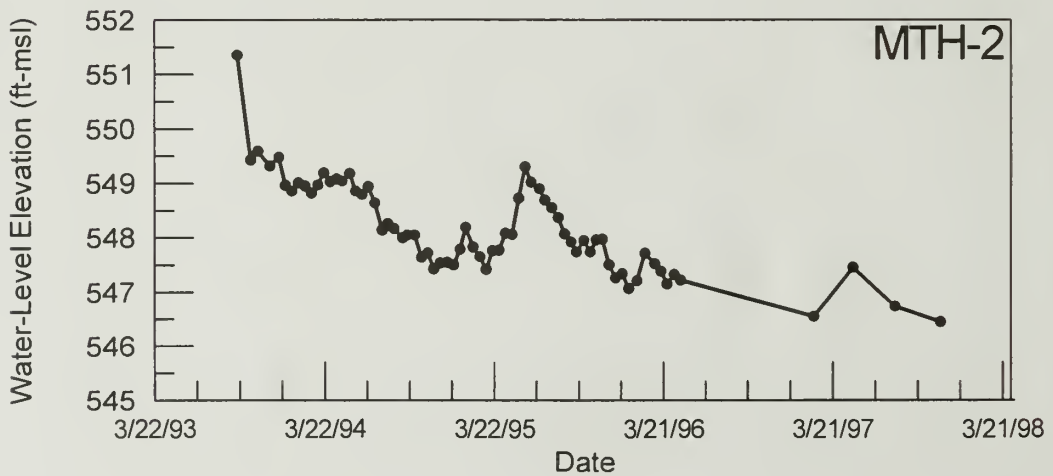
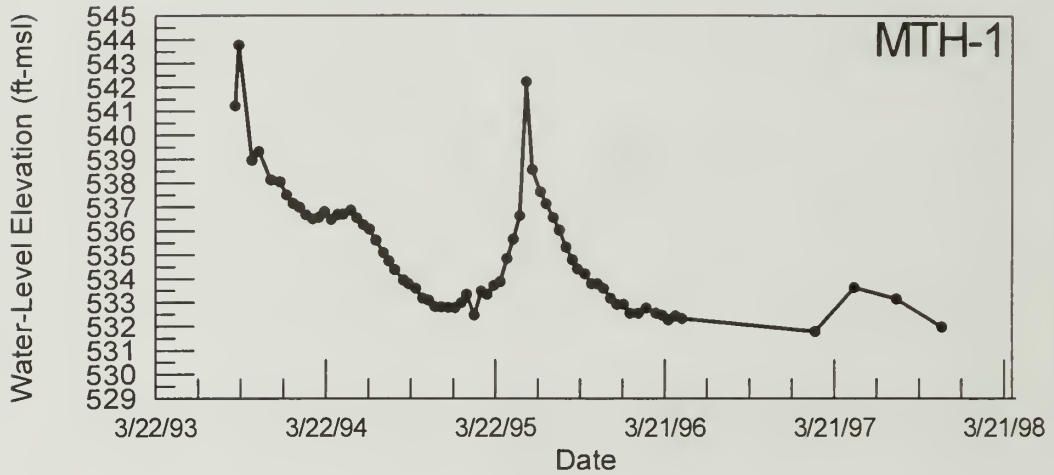
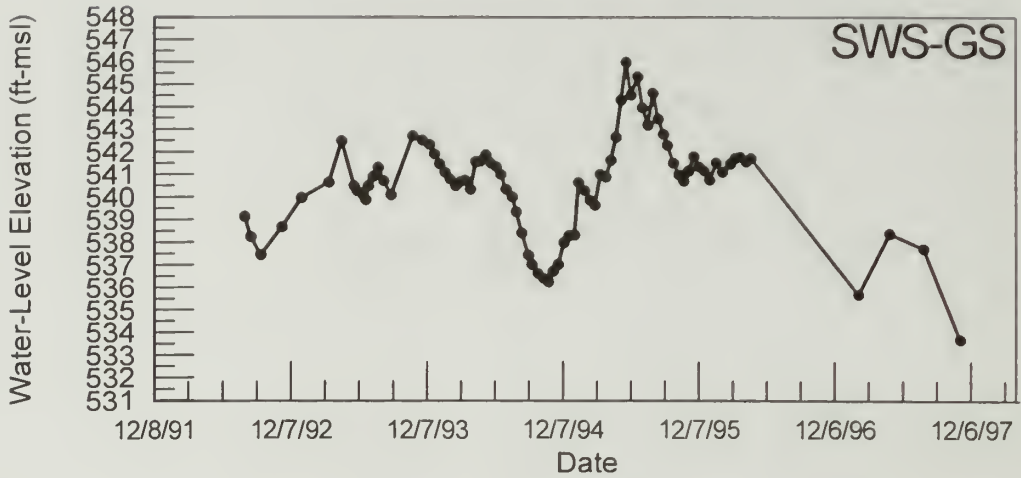
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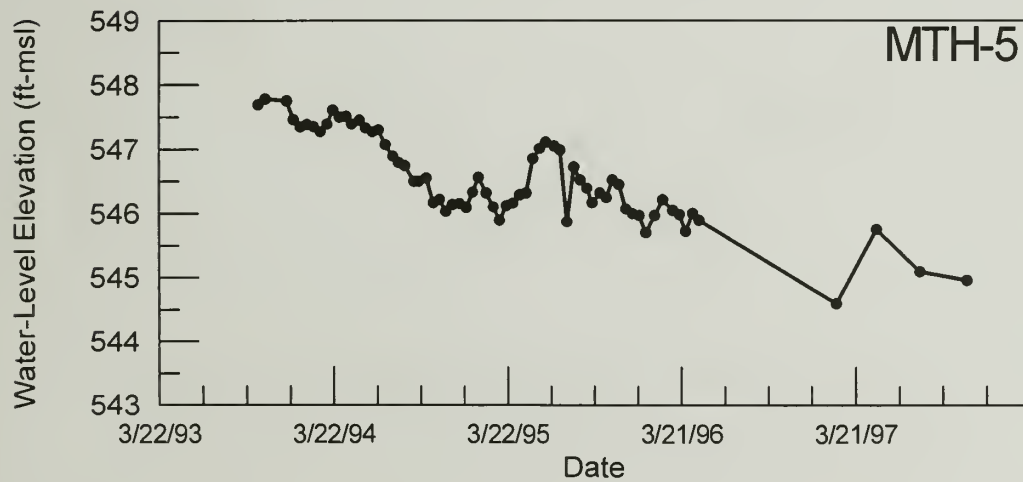
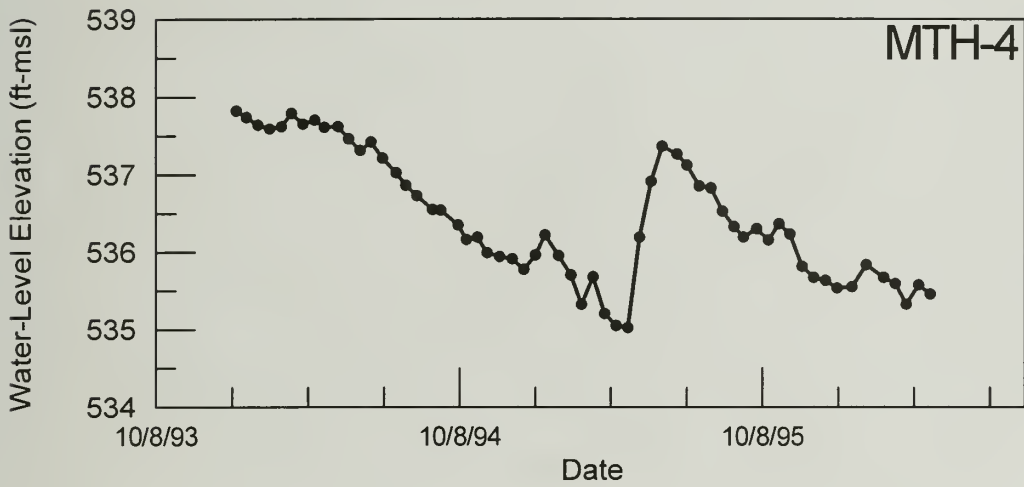
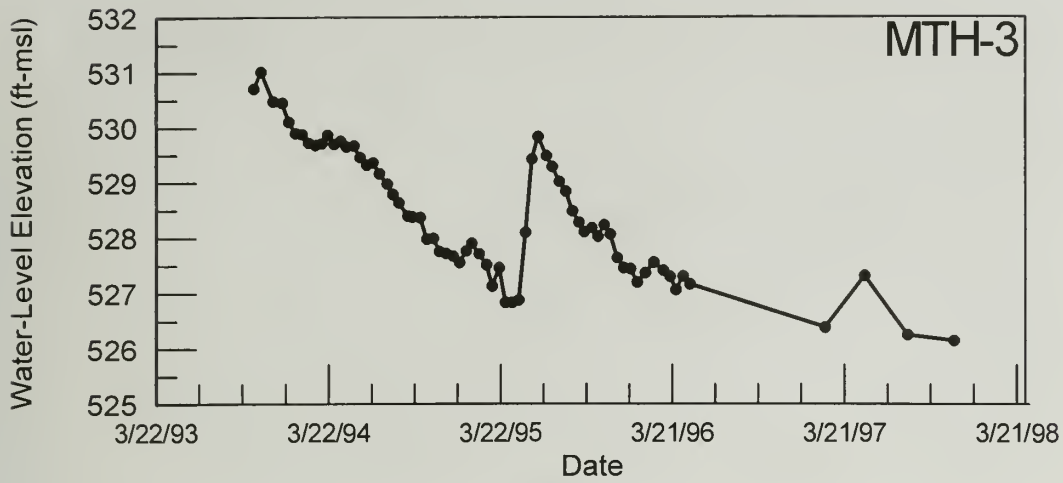
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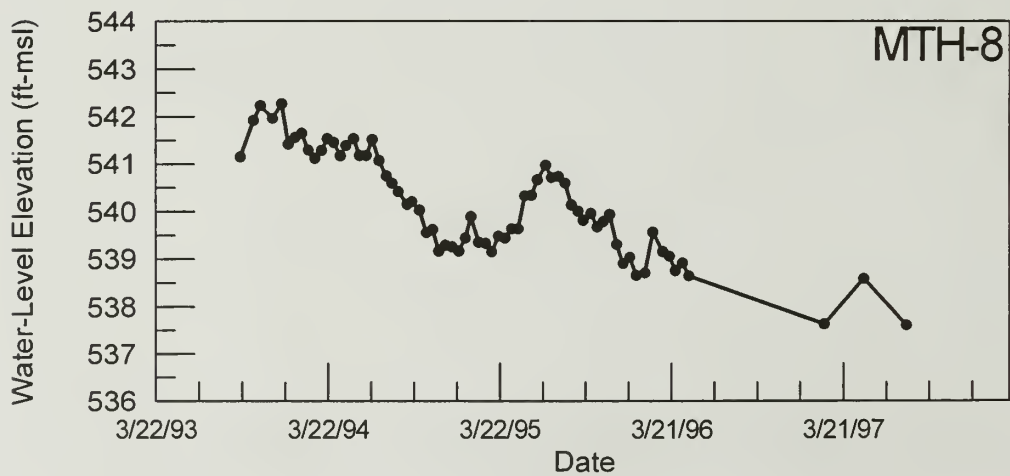
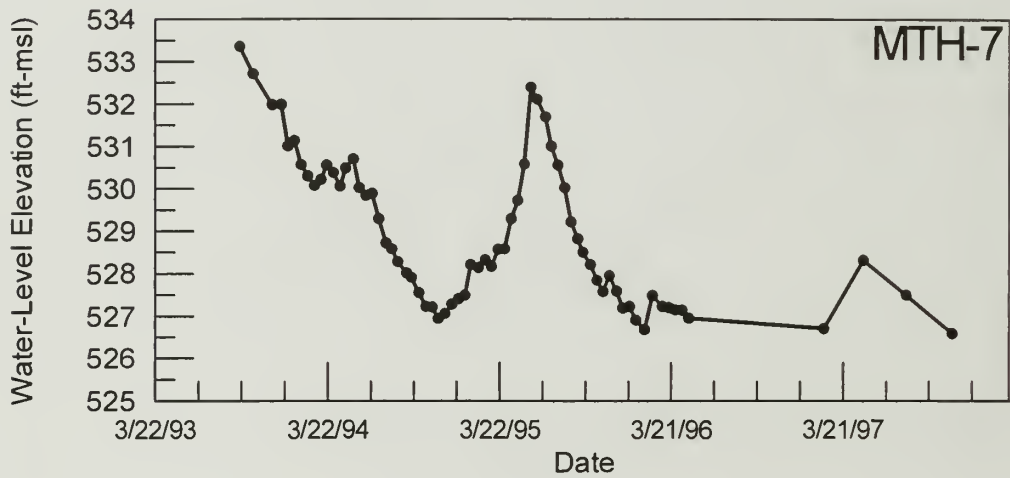
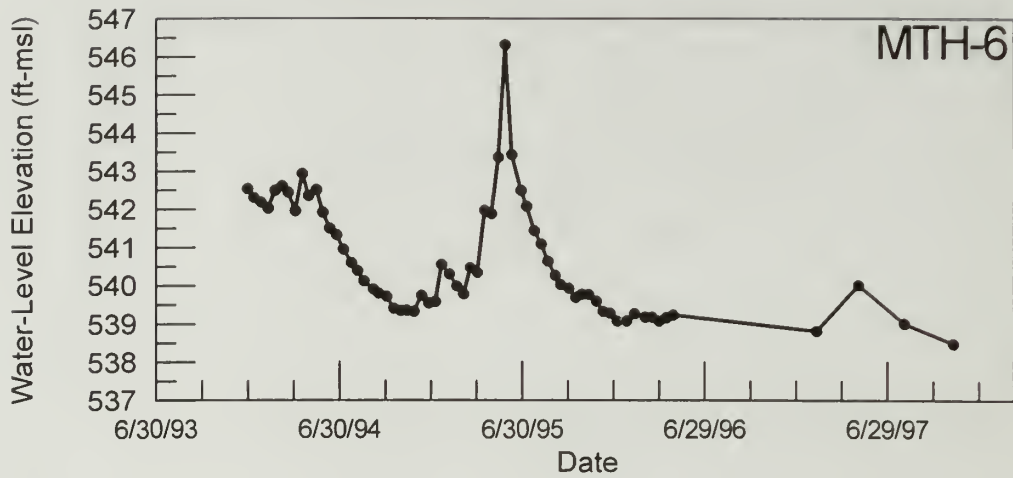
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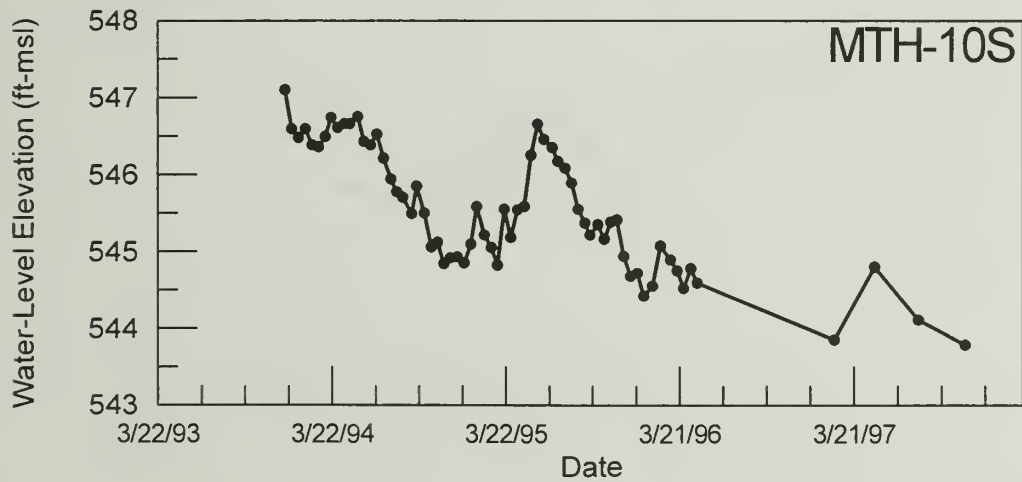
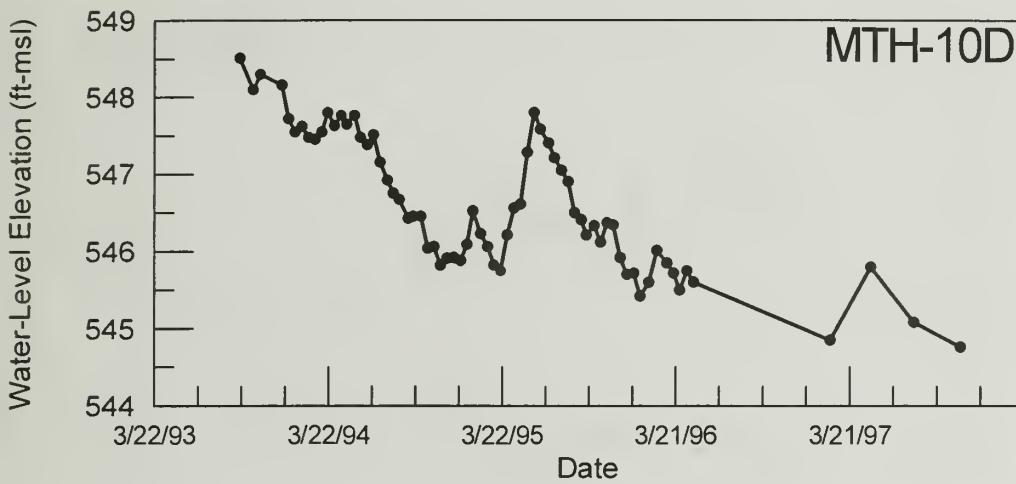
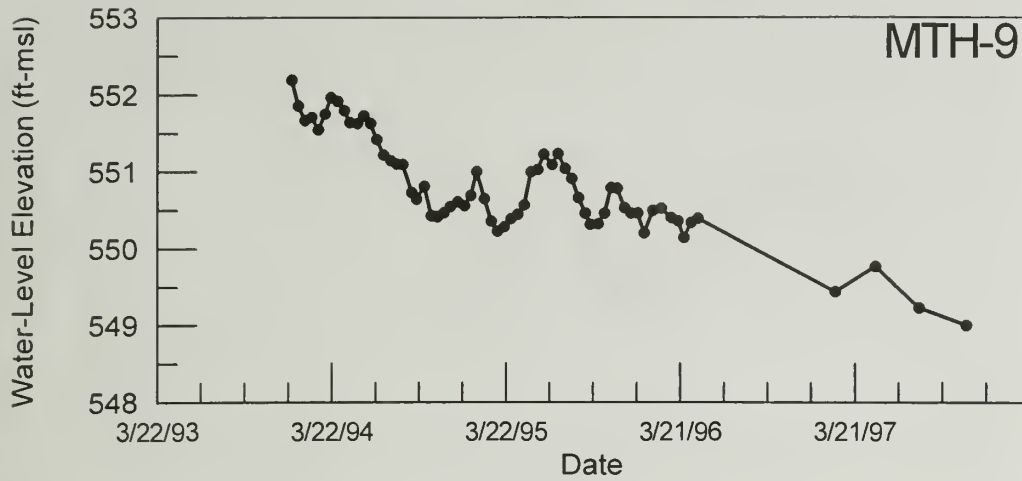
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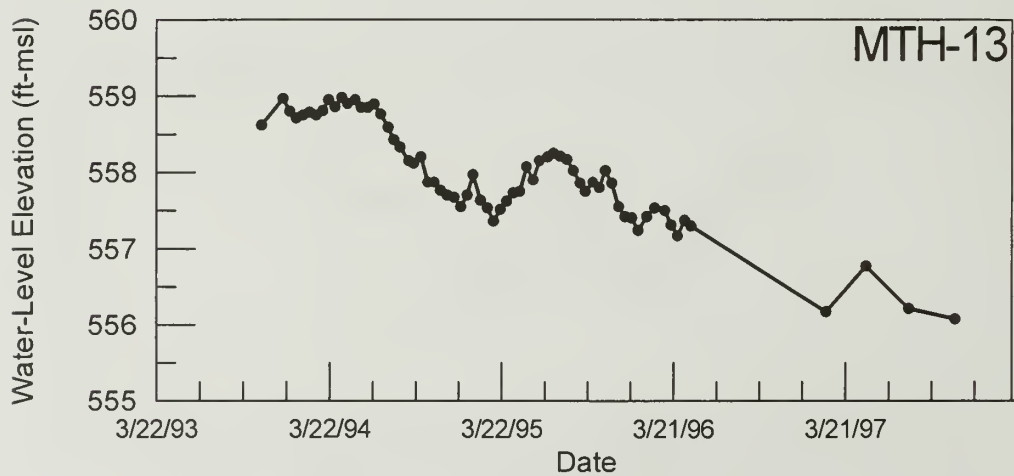
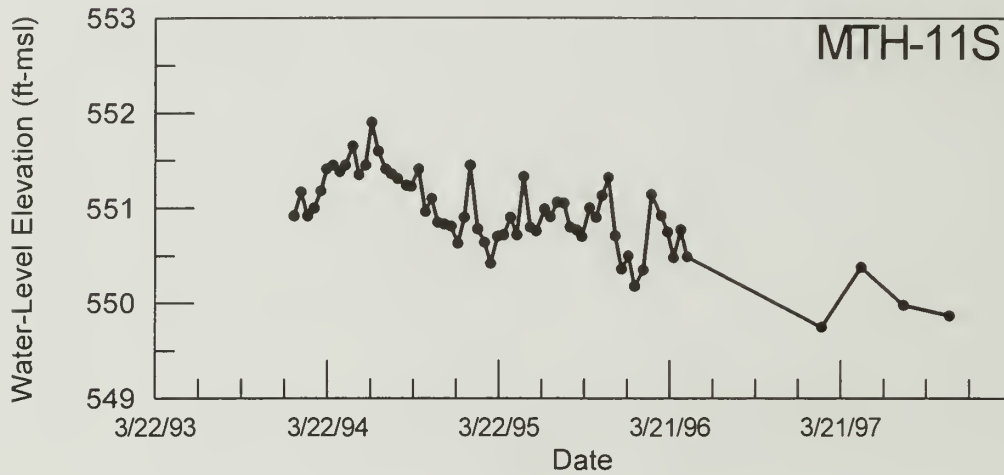
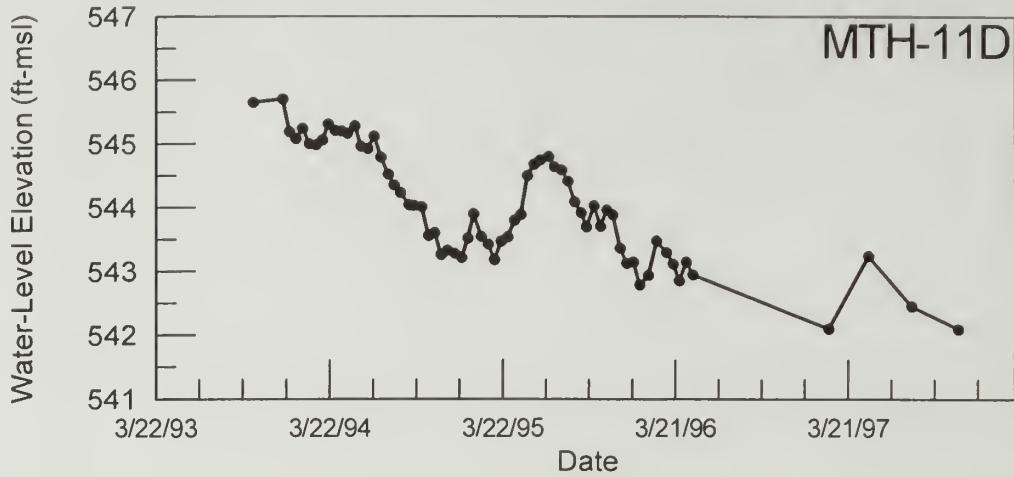
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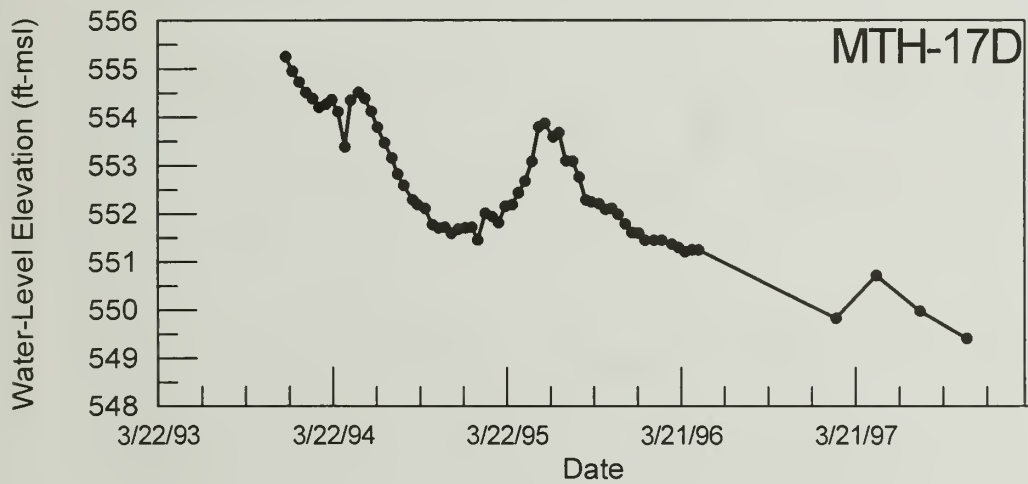
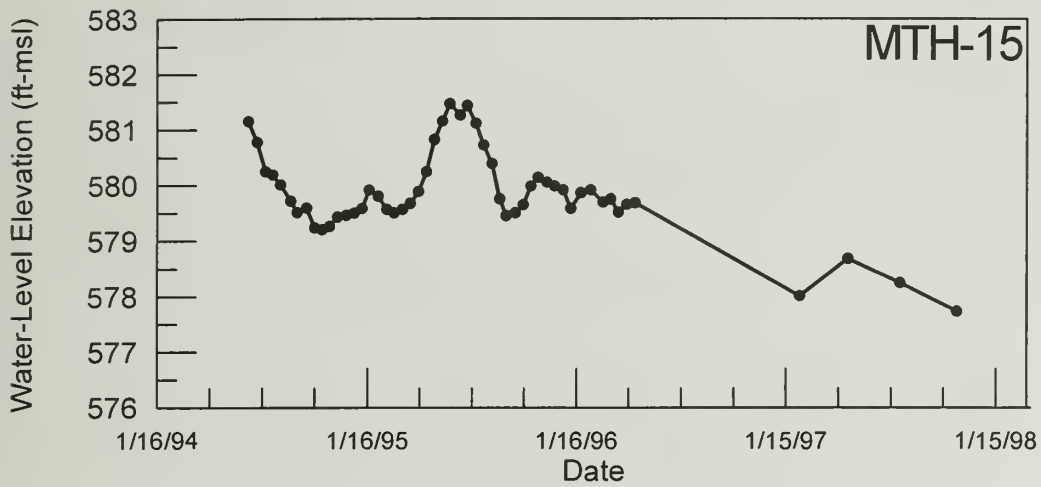
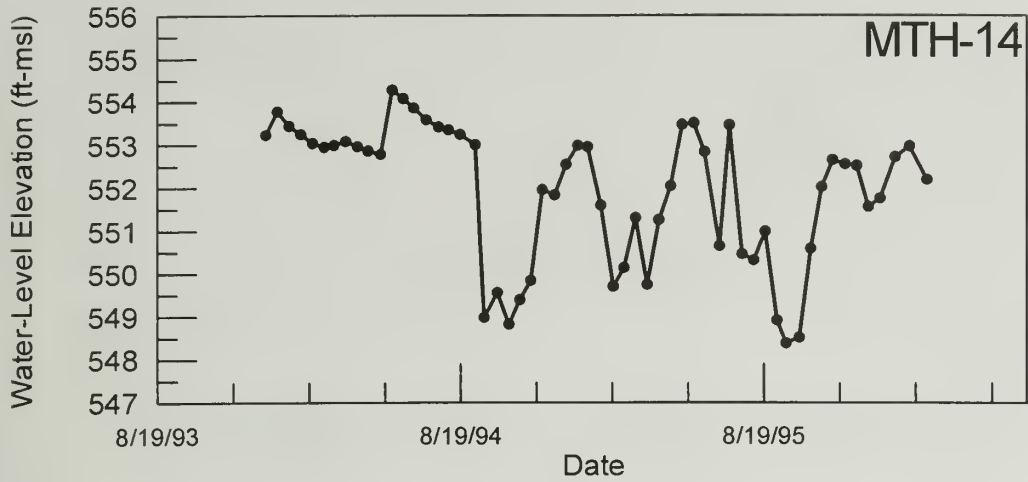
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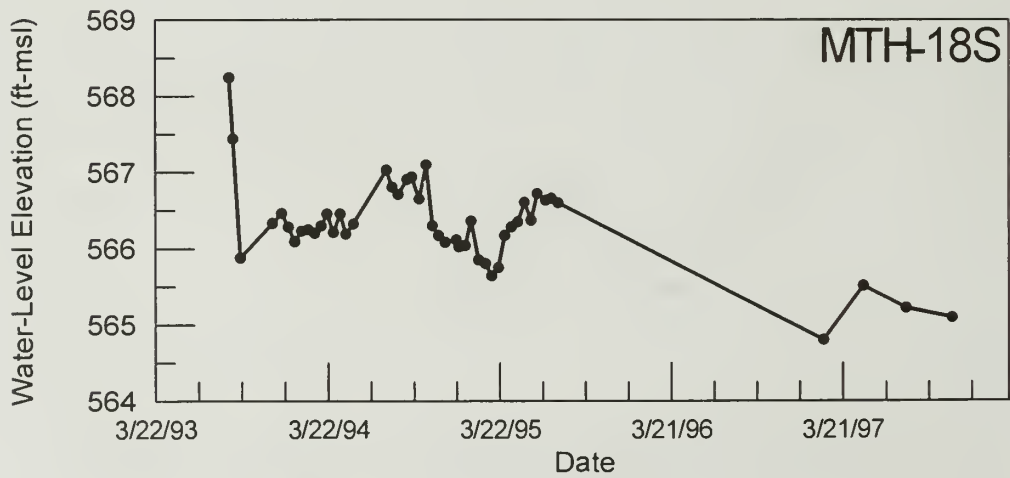
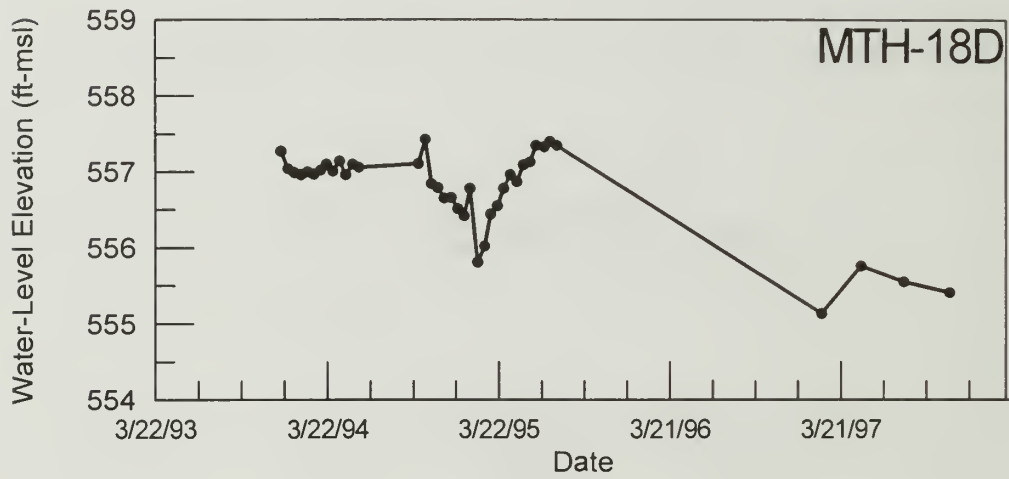
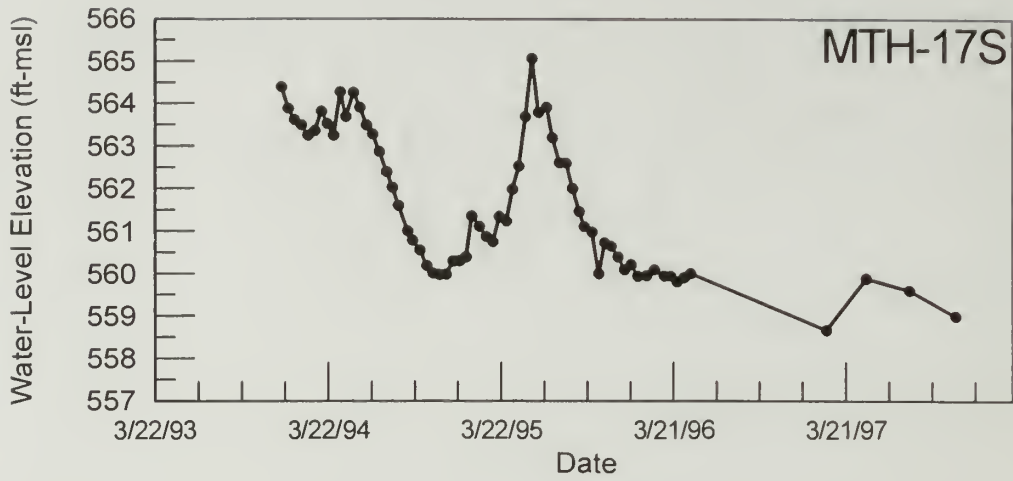
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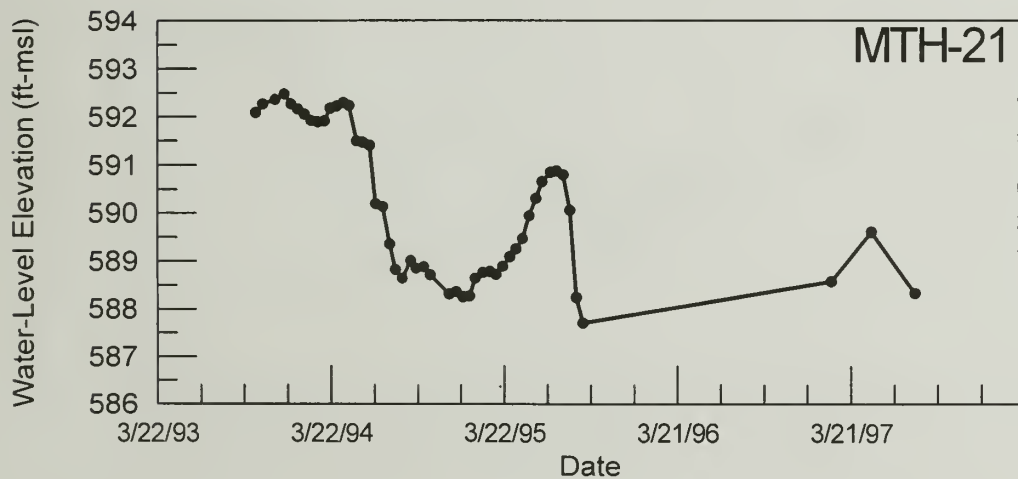
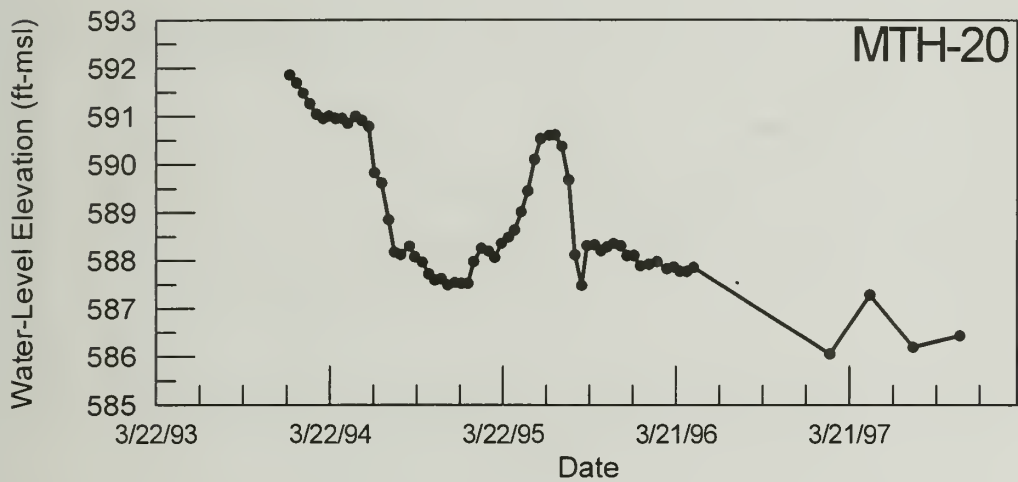
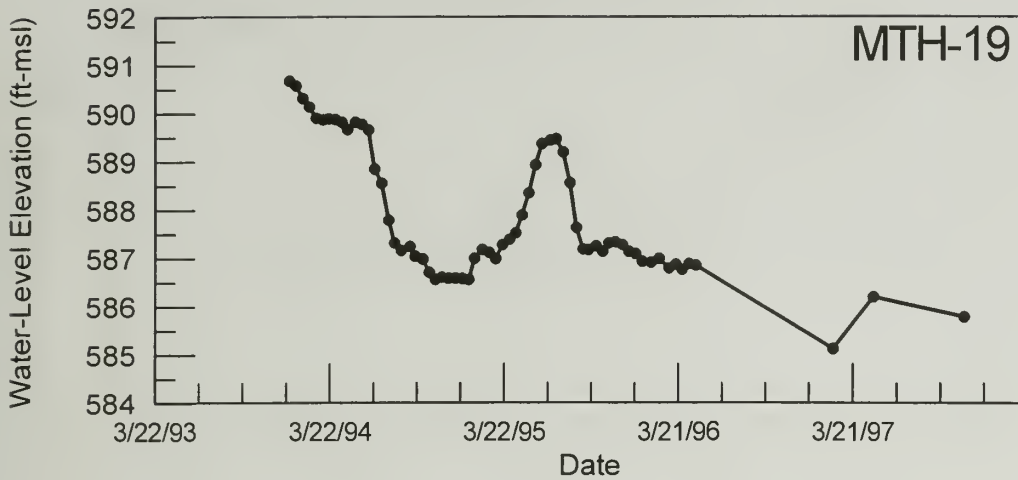
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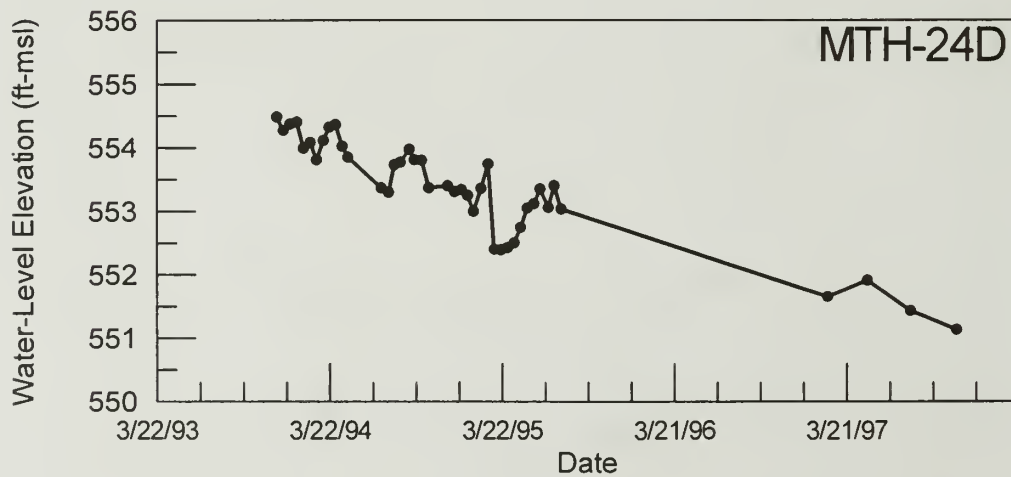
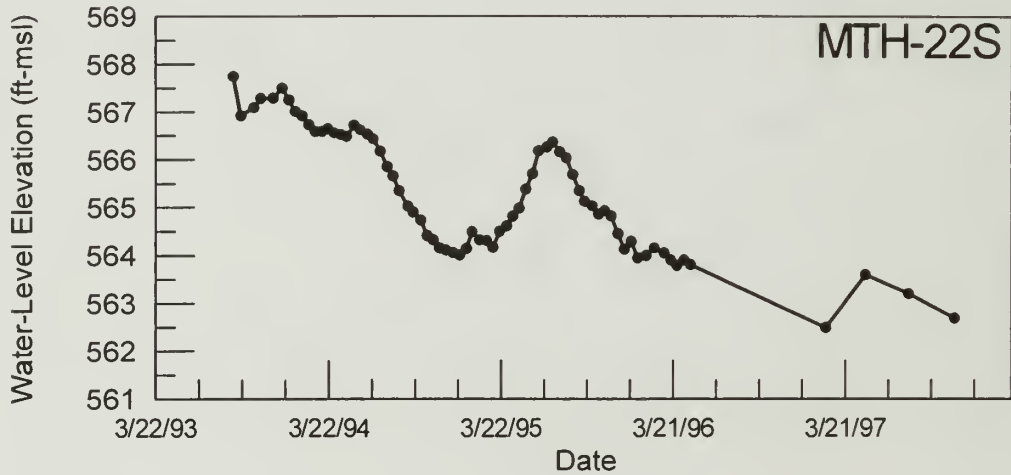
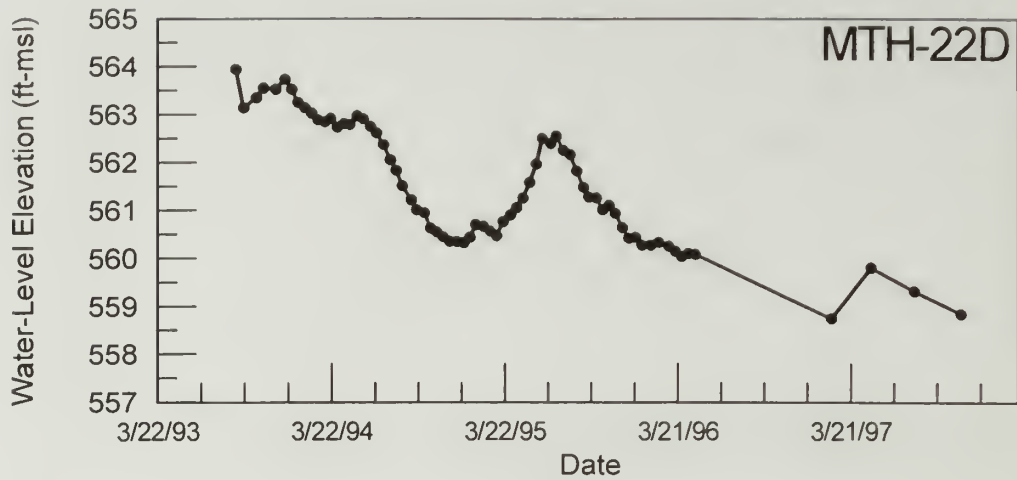
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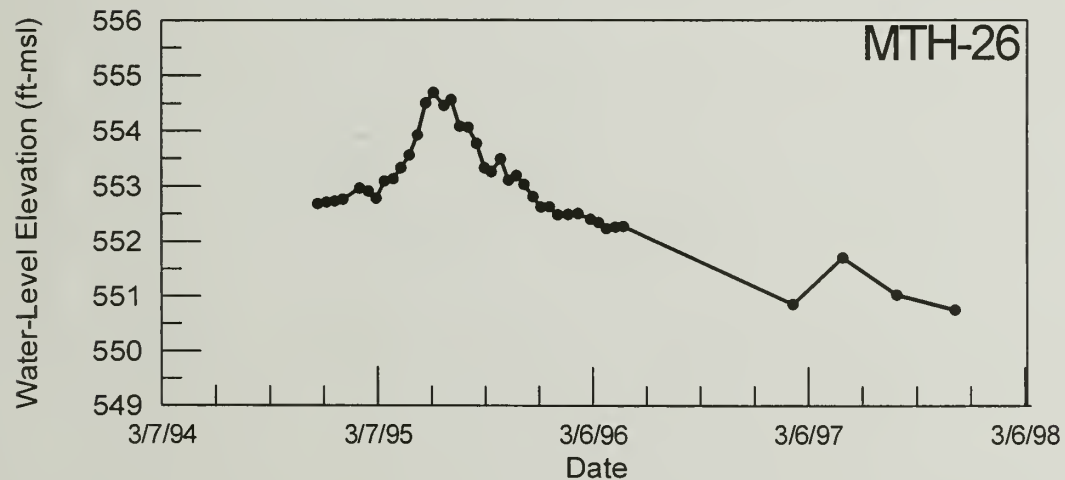
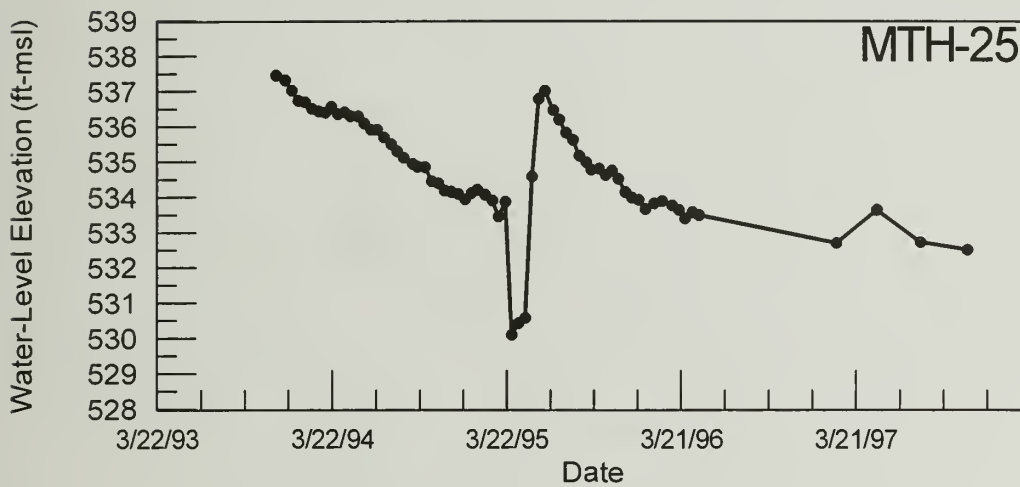
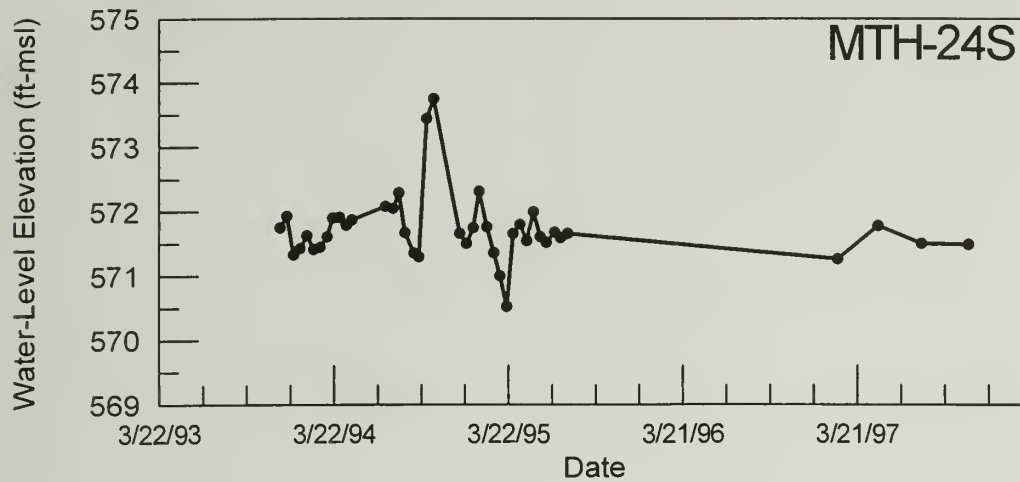
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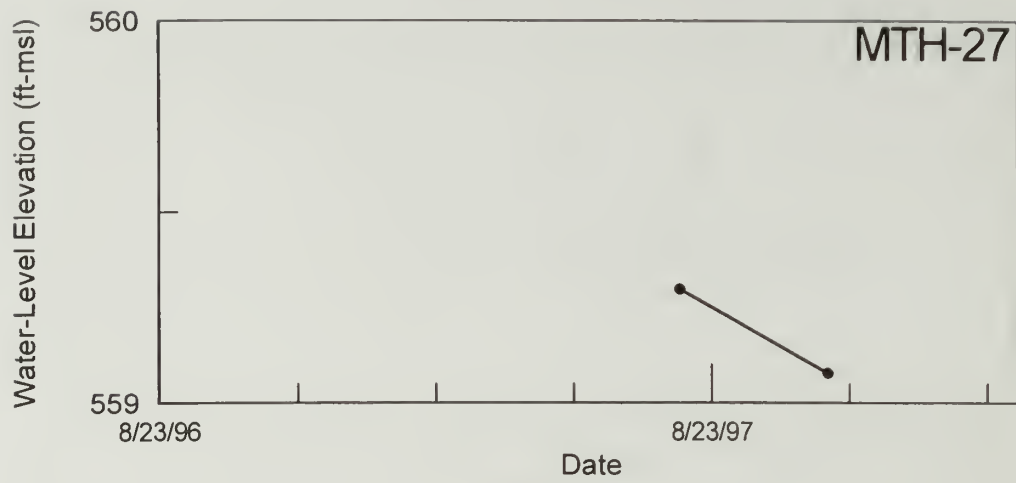
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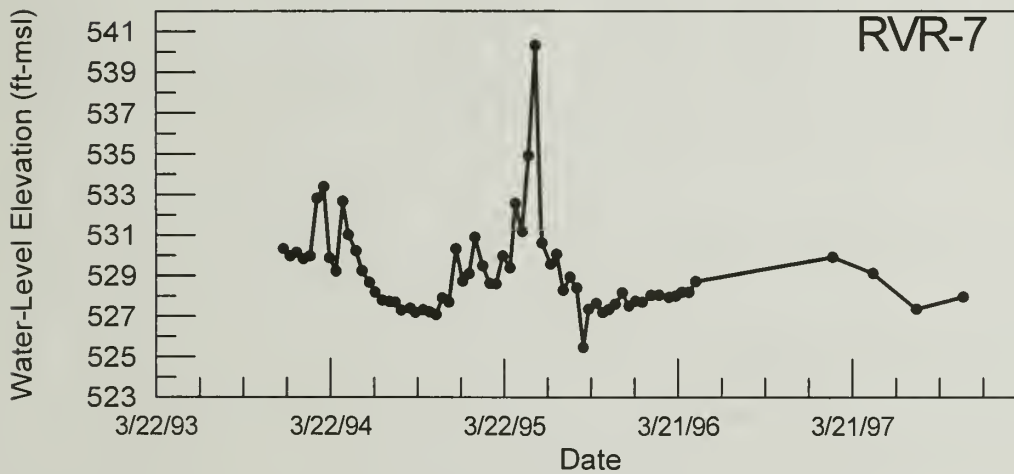
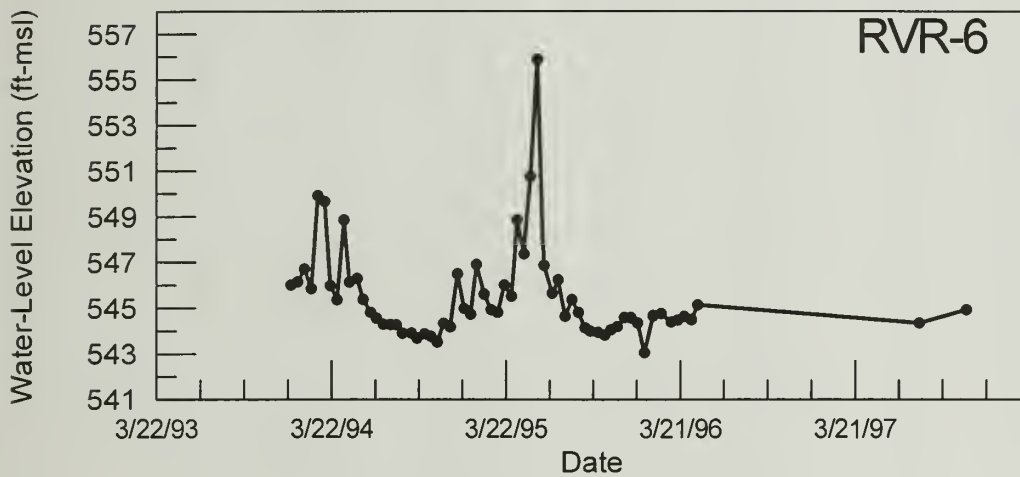
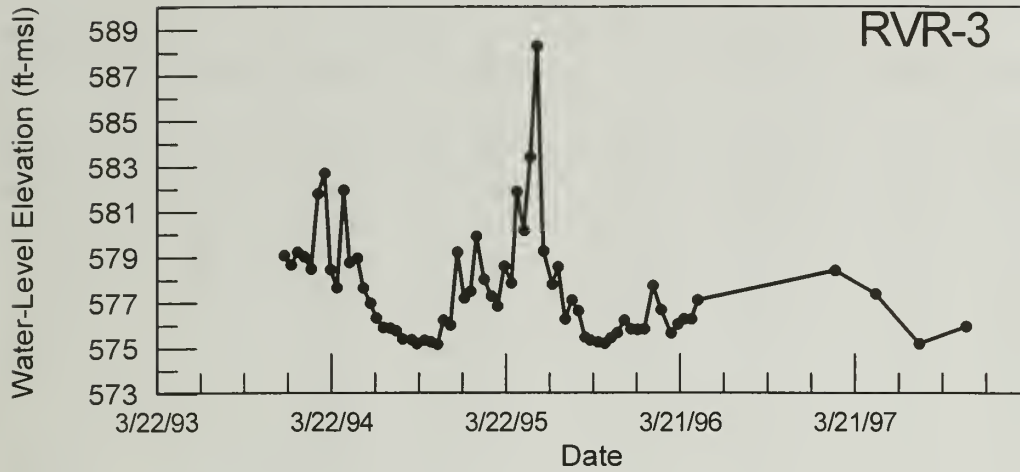
APPENDIX D. CONTINUED



APPENDIX D. CONTINUED



APPENDIX D. CONCLUDED



APPENDIX E. 1994 GROUND-WATER WITHDRAWALS IN THE STUDY AREA

<i>Lambert-x</i>	<i>Lambert-y</i>	<i>Aquifer</i>	<i>Annual pumpage</i>	<i>gpd</i>	<i>Lambert-x</i>	<i>Lambert-y</i>	<i>Aquifer</i>	<i>Annual pumpage</i>	<i>gpd</i>
3065810	2765680	GL	2925360	8015	3003266	2727658	SM	55766000	152784
3074909	2633197	GL	9927560	27199	3003216	2727457	SM	18591000	50934
3035424	2731918	GL	2800000	7671	3007652	2758982	SM	87891000	240797
3022113	2729582	GL	5838000	15995	3007541	2758701	SM	82483000	225981
3075101	2633450	GL	9927560	27199	3007475	2758306	SM	100061000	274140
3035350	2731842	GL	2800000	7671	3089992	2730044	SM	20800000	56986
3075187	2631703	GL	9927560	27199	3089995	2729964	SM	12000000	32877
3001156	2776743	GUB	6144072	16833	3007830	2758901	SM	165242000	452718
3101743	2749061	GUB	6525400	17878	3007839	2758457	SM	85470000	234164
3022154	2727383	GUB	1100000	3014	3007694	2758451	SM	129154000	353847
3073573	2632877	GUB	9927560	27199	3009960	2759929	SM	96584000	264614
3104554	2625536	GUB	4800000	13151	3091915	2652853	SM	16964500	46478
3073574	2632824	GUB	9927560	27199	3092234	2653211	SM	16964500	46478
3101856	2749008	GUB	6525400	17878	3010060	2759933	SM	128029000	350764
3022719	2729804	GUB	2502000	6855	3051547	2695631	SM	27027100	74047
3038807	2628380	GUB	730000	2000	3051546	2695661	SM	11941000	32715
3031541	2600947	GUB	85430000	234055	3035398	2731942	SM	60000000	164384
3032075	2600942	GUB	19310000	52904	3024278	2691367	SM	15239000	41751
3016464	2629290	SM	1314000	3600	3023971	2691200	SM	15239000	41751
3104048	2625857	SM	5551000	15208	2987299	2673691	SM	51166940	140183
3112827	2594190	SM	300630	824	2987269	2673681	SM	15577848	42679
3016461	2629447	SM	3358000	9200	3052064	2661808	SM	10009000	27422
2977060	2605402	SM	1623050	4447	2976890	2659649	SM	190600	522
2976454	2606089	SM	3807225	10431	3024046	2768907	SM	12621400	34579
2977200	2606498	SM	3807225	10431	3022460	2767224	SM	29255300	80152
3085713	2709568	SM	202342200	554362	3096259	2715132	SM	164810200	451535
3004233	2646511	SM	14957400	40979	3090010	2715224	SM	136112300	372910
3112727	2594296	SM	1517700	4158	3093094	2715281	SM	225068100	616625
3015247	2631671	SM	4818000	13200	3069285	2738293	SM	2943360	8064
2977060	2605396	SM	1623050	4447	3076790	2696348	SM	11132500	30500
3004234	2646461	SM	16841400	46141	3020782	2772099	SM	1000000	2740
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3060503	2766467	SM	12000000	32877	3043508	2731292	SM	1950000	5342
3061893	2766687	SM	7800000	21370	3001158	2776243	SM	1496748	4101
3066466	2765690	SM	216600	593	3046559	2765881	SM	15470300	42384
3020337	2772438	SM	4000000	10959					

Notes:

Annual pumpage - total 1994 pumpage, in gallons

GL - Glasford aquifer

gpd - gallons per day

GUB - lower Glasford/upper Banner aquifer

Lambert-x - east-west Lambert coordinate, in feet

Lambert-y - north-south Lambert coordinate, in feet

SM - Sankoty-Mahomet Sand aquifer

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